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22nd September 1944.

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M.R.P. 7
5 February 1942

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THE MEASUREMENT OF HUMIDITY AT LOW TEMPERATURES

Explanatory Note: By E. Glückauf

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The attached report by Dr. E. Glückauf is the second of his quarterly reports on the measurement of humidity at low temperatures. It deals with experiments to determine the characteristics of the electric resistance hygrometer developed by the National Bureau of Standards, U.S.A. Briefly the hygrometer unit consists of a tube coated with a synthetic resin impregnated with lithium chloride and wound with two coils of wire. The resistance between the coils is a measure of the relative humidity.

The scope of the experiments had been dictated by the need for a hygrometer for use on aircraft up to stratosphere levels. The previous report which dealt mainly with goldbeaters' skin as a humidity element but included preliminary experiments on the electric resistance type, indicated that the lag of these instruments at low temperatures is largely determined by the rate of air flow past the elements.

Quarterly Report on work done for the Meteorological Office during October to December 1941 at the Science Laboratories of the Durham Colleges, Durham

The apparatus for calibrating hygrometers was adapted for use with larger quantities of air and higher velocities. The arrangement of the low temperature plant for hygrometer calibration is shown in figure 1. Air from a compressor enters under a pressure of 130-140 atm a steel cylinder, which is filled with silica gel. The silica gel has been prepared by heating for one hour at 180°C and, then, has been cooled down in a vacuum of 0.1 mm Hg. The present capacity of the steel cylinder is 20 cbf, but it is being replaced by a 100 cbf cylinder to ensure better drying.

After passing through the drying cylinder the air is expanded in a reducing valve from 130 to about 2 atm, and then passes through the "distribution stopcock", which divides the air current into two, in a definite ratio, one of which is to be saturated with respect to ice and afterwards reunited with the rest of the air.

The distribution stopcock is shown in figure 2. The actual stopcock (Part 2, figure 2) consists of a hollow cone with 9 equal holes of about 1 mm diameter, placed at 20° to each other, and leading into the centre of the stopcock. From Part 1 of figure 2 it can be seen, that, if in position, one of the holes will always be covered, leaving 8 through which the air can pass. By adjusting the position of the stopcock, nine different air flow ratios can be obtained, without altering the total amount of air passing. In this way the following relative humidities (with respect to ice) can be obtained: 0, 12.5, 25, 37.5, 50, 62.5, 75, 87.5, 100%, which cover a range sufficient for calibrating purposes.

(In the experiments described in the reports for the months July-September a different arrangement using several glass stopcocks was used, which was based on the same principle, but gave a greater number of combinations. It could, however, not be used for large quantities of air on account of the much higher pressures involved.)

To guarantee the correct working of the distribution stopcock it was necessary to ensure equal pressure on the two outlet sides. Friction in the coils of the cooling tank caused considerable pressure differences if the quantities of air passing through the two coils were much different (e.g. in the cases of 12.5 and 87.5% relative humidity). To avoid errors of this kind additional flow resistance could be provided by two valves C and D, one of which was adjusted until the mercury manometer showed no difference of pressure.

The two air currents passing through the valves C and D were cooled down to the temperature of the bath by passing through two helical coils of tubing of 5/8" and ½" diameter respectively. The wider coil had a ribbon of cotton bandages

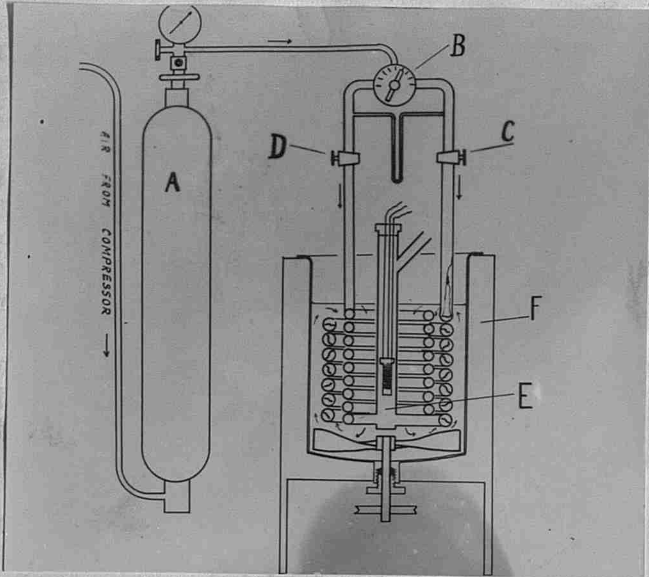

B
D
C
A
AIR FROM COMPRESSOR
F
E

Figure 1. Low temperature testing plant.
A. Drying vessel filled with silica gel.
B. Distribution stopcock(see figure 2).
C.D. Valves for regulating flow resistance to air.
E. Testing chamber for hygrometers.
F. Thermo-insulation for low temperature bath.

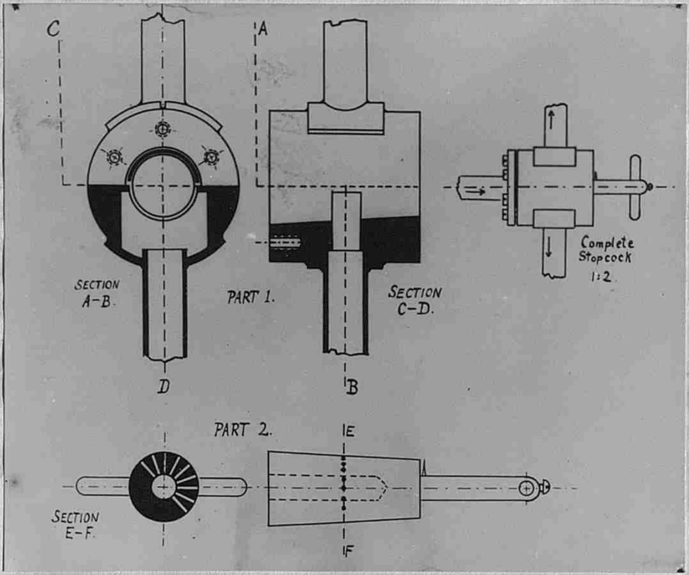

C
A
SECTION A-B.
PART 1.
SECTION C-D.
Complete Stopcock 1:2
D
B
PART 2.
E
SECTION E-F.
F

Figure 2. Distribution stopcock, used for obtaining air of known relative humidity.
(Sections 1:2.8 nat.size, complete view 1:5.6 NS.)

held at the ends by copper wire and stretched through nearly the entire length of the tubing (about 15'), which was occasionally moistened with distilled water; this arrangement proved sufficient to saturate the air completely, even at the highest speeds, with respect to ice at temperatures below -10°. This method proved superior to the freezing out of surplus water vapour from moist air, as it avoided the formation of snow inside the tube, which at the high air speeds used, would have been driven right through the apparatus.

(In the case of higher temperatures, however, the air was moistened by passing through a wash bottle filled with a suitable solution, so as to avoid that the wet ribbon lost its water too quickly.)

The two coils, one containing dry, the other containing saturated air, were connected excentrically to the bottom of a tube of 1" internal diameter, which served as testing chamber. The excentrical connection is of importance, as otherwise the two air streams do not mix efficiently, especially at very high and very low humidities, where the two velocities are rather different.

The low temperature bath was filled with trichloroethylene, which, though its vapours were disagreeable, had the advantage of being non-inflamable, keeping its low viscosity at 170°, and producing relatively little foam (as compared with ether or pentane). The low temperature was produced by constantly adding small pieces of solid CO_2.

To obtain a uniform temperature, the liquid was vigorously agitated by a motor driven stirrer, the shaft of which passed through a stuffing gland in the bottom of the vessel. The blades were bent in such a way as to produce a fast circulation of liquid in the direction indicated in figure 1. The bath container was held in a wooden frame and was only moderately insulated (F), as the transfer of heat through the insulation was negligible compared with the heat introduced by the passing air.

The units to be tested were either inserted directly into the testing chamber or, if very high air flows were desired, were held in ebonite holders which narrowed down the cross section of the air flow, and thus increased flow and turbulence. In the experiments giving air speeds of about 100 miles-hour, the cross section of air flow was narrowed down to 1 sq.cm.

Among experiments carried out were: (1) calibration of several units with 9%, 6% and 3% LiCl (20 turns). (2) calibration of several 5% LiCl- 5 turns units. (3) Combining several units for the purpose of obtaining a hygrometer with indications independent of temperature. (4) Building and calibrating a temperature-independent hygrometer (embodying several resistances in one unit).

The equilibrium values for the resistances at various temperatures and humidities of a 3% LiCl-20 turns unit and of a 6% LiCl-20 turns unit are given in figures 3 and 4 respectively. Units of the same class show the same type of arrangement for the calibration lines, but the absolute resistances of the individual unit may differ considerably (to one order of magnitude).

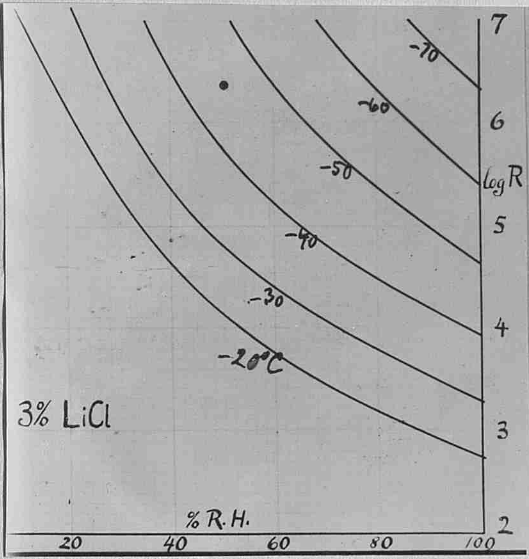


Figure 3

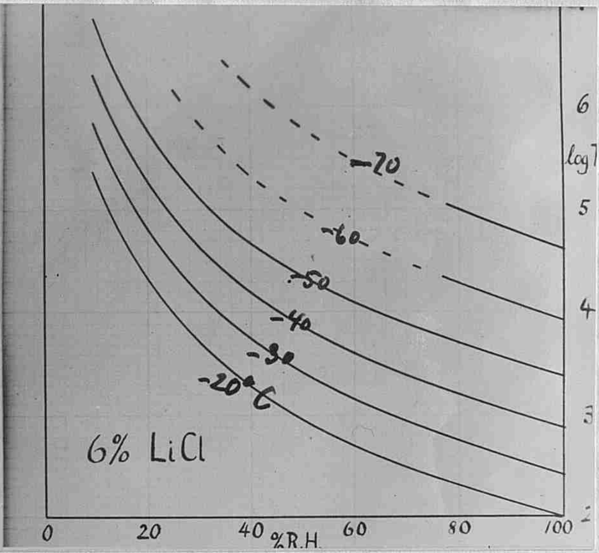


Figure 4

Resistance of a 3% and of a 6% LiCl unit as function of temperature and relative humidity.

The resistance of various units at saturation as function of the temperature is shown in figure 5. It is seen that the effect of temperature is least marked in the case of high LiCl-contents, but the same is the case for the effect also of relative humidity on resistance. The resistances of the 5-turns elements follow curves similar to those of the 20 turns elements, but the absolute values are more than 100 times higher.

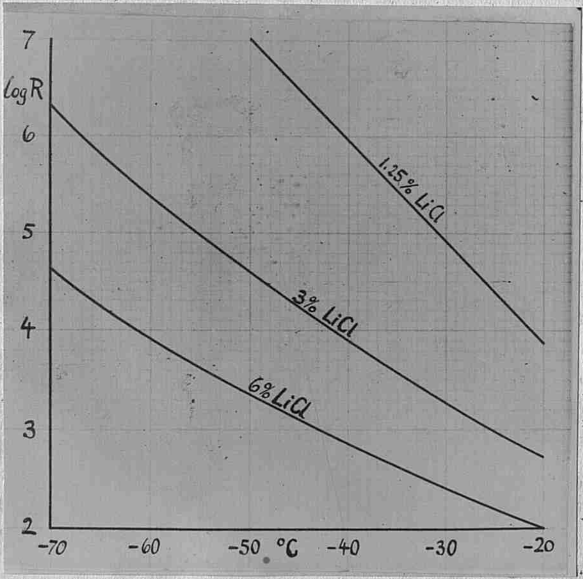
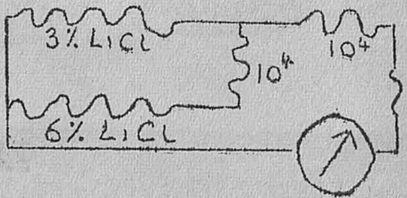


Figure 5

Effect of temperature on resistance for different percentages of LiCl. (20 turns units, saturated).

The resistance ratio R_H/R_{sat} was found to be constant at all temperatures between -20 and -50° for the 6% LiCl elements, which offers possibilities for the construction of a temperature independent hygrometer. For lower percentages of LiCl R_H/R_{sat} is far from constant, for different temperatures.

While the resistance of the 6% LiCl unit is not as sensitive to changes of relative humidity and temperature as the units with lower LiCl content, it still varies over five powers of ten for a change in humidity from 10% at -50° to 100% at -20°C. This is too much to be measured directly by electrical instruments, but a useful arrangement, measuring nearly the full humidity range from -20° to -60° can be obtained by using an arrangement like e.g. :



or, giving a more proportional scale, by the arrangement shown in figure 6, which also represents the calibration curves of this arrangement.

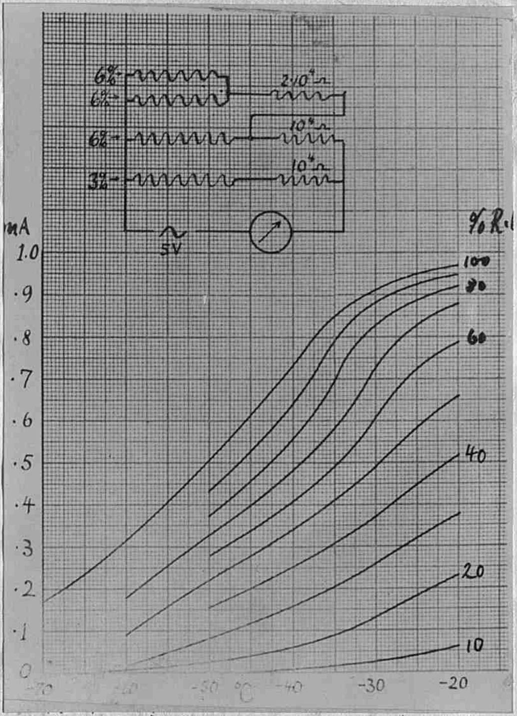


Figure 6

Arrangement and calibration chart of low temperature hygrometer.

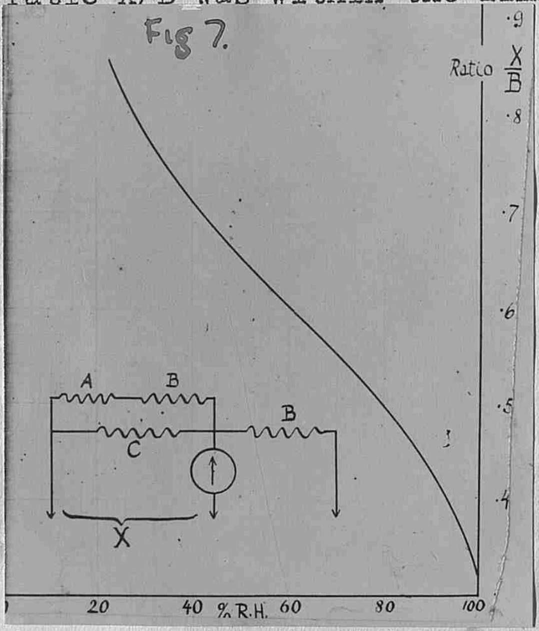
A disadvantage, however, of these arrangements, from the practical point of view, is that the readings of the micro-ammeter are a function not only of the humidity but also of the temperature, so that without a complete calibration chart and knowledge of the temperature, the humidity cannot be obtained. This disadvantage which is also inherent in all combinations described in the N.B.S. Research papers Nos. 1265 and 1329, can be avoided by using elements which employ the 6% LiCl solution only, but are made up by using different lengths and spacings between the wires. The temperature effect can then be eliminated by measuring not the absolute resistance but the resistance ratio between such a combination and an always 100% saturated element preferably made from the same solution. The arrangement shown in figure 7 was tried and was found, in this respect, very successful. In the diagram

A = 6% LiCl - 20 turns, exposed,
B = 5% LiCl - 5 turns, saturated
C = 5% LiCl - 5 turns, exposed

Tests between -20° and -50°C showed that the resistance ratio X/B was within the limits of the experimental error independent of the temperature. The calibration curve of this unit is shown in the figure (7).

In this case all the units used were standard units which could be purchased ready made in U.S.A., which no doubt has great advantages in many respects. It was thought, however, that all these different units might be combined suitably into one single unit without undue increase in size. After several experimental units had been tried, the design shown in figure 8 proved to be the most satisfactory solution. The hygrometer unit consisted of 4 resistances:

Fig 7.



Arrangement and calibration curve of temperature independent combination.

A. 20 turns at 16 turns per inch
B. 2 turns at 16 turns per inch
C. 1 ring of 1/16 inch width, satur.
D. 1 ring of 1/8 inch width, satur.

coated with 6% LiCl solution

The arrangement of the resistances is shown in the photograph, figure 9 in natural size. The section containing

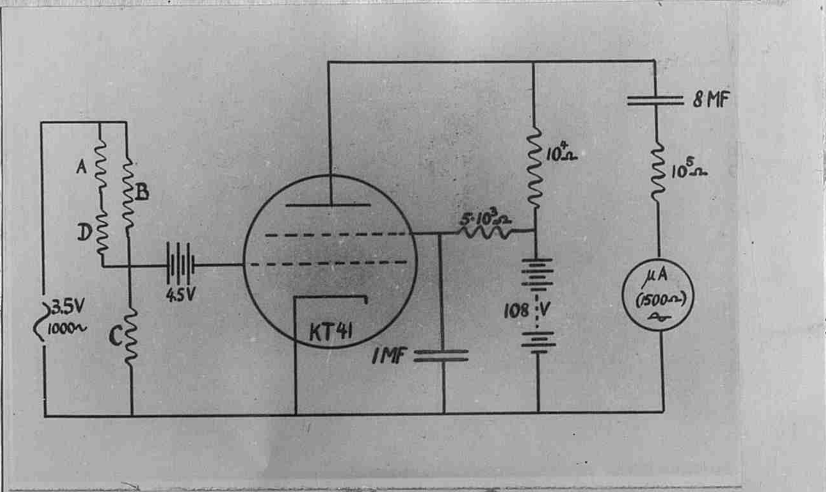

A
B
D
C
3.5V
1000Ω
4.5V
KT41
1MF
5·10³Ω
10⁴Ω
10⁵Ω
8MF
108 V
μA
(1500Ω)

Figure 8.

Circuit for directly indicating hygromete

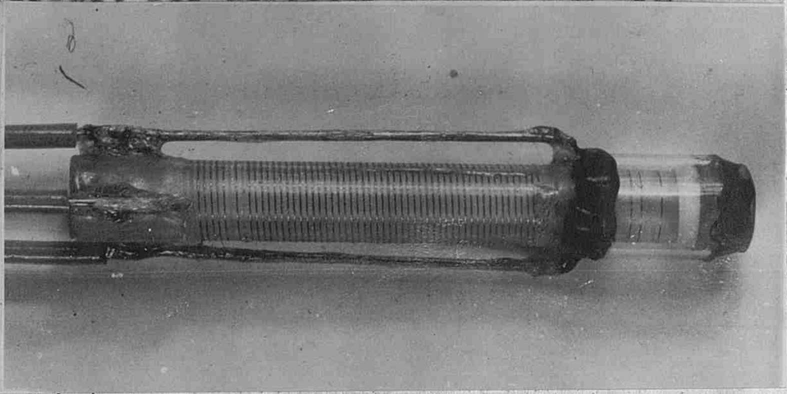
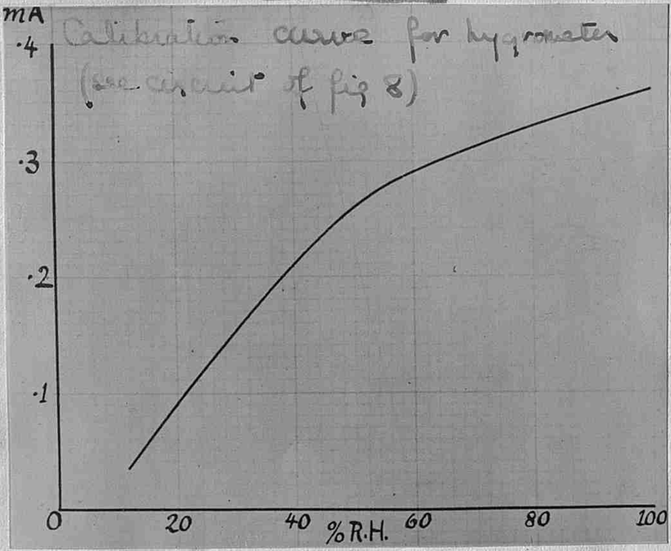


Figure 9

Temperature-independent single-unit hygrometer.

The resistances C and D was covered with glass tubing, the ends of which were sealed with a solution of plystyrene or picein wax, and saturation was produced by a small piece of wet filter paper inside the closed ring space. (It would be preferable to use polystyrene tubing instead of the covering glass tube, as the different expansion coefficients give rise to cracks and leakages). Using the circuit shown in figure 8, the deflection of the micro-ammeter were found to be practically independent of the temperature between -16°C and -53°C. The calibration curve of this arrangement is given in figure 10.



It suffers rather from lack of proportionality, but this is being rectified by decreasing the resistance of C and by increasing the grid bias of the valve.

Experiments showed that the lag of the 6% LiCl unit (20 turns) at temperatures below -50°C was rather less than expected from the extrapolation of the values at higher temperatures. This is very difficult to explain, but a similar feature was found also in earlier experiments with the hair hygrometer. The following table gives the average lag factors λ (according to

$$\ln \frac{H_\infty - H_0}{H_\infty - H} = \frac{t}{\lambda}$$

if an air speed of approximately 100 miles/hour is used (600 cbf of air per hour passing the unit parallel to its axis in a layer of about 3 mm thickness).

T°C	λ (minutes)
-20	0.1
-30	0.3
-40	1
-50	3
-60	7
-70	12

It is not unlikely that at the lower pressures of the stratosphere the values of λ may be reduced further, which would bring the lag to reasonable figures even at the lowest temperatures to be expected.

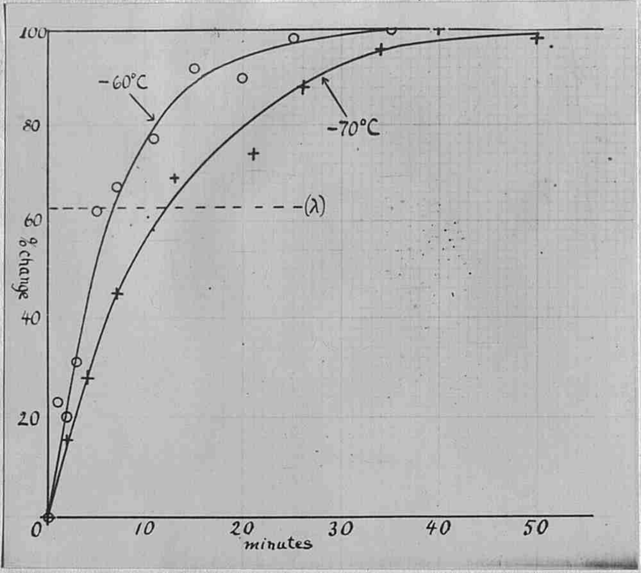


Figure 11

Lag curves at -60° and -70°C.

Figure 11 shows the lag curves measured for a sudden change from 50% to 75% relative humidity at -70°C, and from 25% to 37.5% at -60°C, both for a 6% LiCl 20 turns unit. The fluctuations in the measured points arise mainly from the difficulty of maintaining the temperature sufficiently constant in a bath of small dimensions. (A larger testing tank is under construction).

Experiments at present centre on the improvement of the design shown in figure 8, i.e., in the construction of a useful hygrometer, the indications of which are independent of temperature.

(Sgd.) E. Glückauf

January, 1942

Meteorological Research Committee

M.R.P.12.

3 March 1942

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Interim Report on Classification of Synoptic Charts, Persistence and Recurrence of Anticyclones.

By A.H.R.Goldie.

(see Appendix for geog. regions)

At the Conference on long range forecasting, held on 26th June 1941, it was decided, amongst other things, that an examination should be made of cases of persistent anticyclones and the conditions leading to their formation and decay. The preliminary classification of weather charts for this and other purposes was done by Miss F.M.Jones in intervals from other duties. The Northern Hemisphere Charts of the D.W.R. as issued from March 1930 onward were used.

The classification of weather types adopted was based on a digest of results of earlier classifications, namely those of Koppen and van Bebber, Gold and Newnham; and though not quite exhaustive was intended to avoid recourse to the almost endless number of types which would be required to cover every possible case. In practice about 2½% of weather charts could not be fitted into it. The classification is given in an Appendix. Besides analysing 11 years of weather charts in this way various statistical results on frequency, mean duration, persistence and sequence have been got out, as well as daily values of certain quantities suitable for studying atmospheric circulation over the Eastern Atlantic.

The Azores High is predominant, (Type I) with or without extensions, on nearly two-thirds of the days of the year. The persistence of anticyclones in other regions namely in the latitude of the British Isles (50°-60°)(Type II) and in lat. 60°-70° (Type III) has been examined. The occasions where Types II or III persisted for 5 days or more, being rather numerous, have been analysed so far only statistically as to preceding and succeeding types. For example, cases of Type II of this duration were in over 70% of occasions preceded by Type I, but less persistent cases of Type II are about equally often preceded by Type I.

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Type III however is not often a further development from Type II, but mostly develops from and relapses into Type IB i.e. the type where, in addition to an anticyclone in the Azores region, there is high pressure also in the Greenland region.

An examination of individual cases where Type II was more persistent and lasted for 10 days or more shows that in 8 of the 9 summer cases and in 6 of the 10 winter cases it developed from the Azores High or arose actually as a northward displacement of the Azores High. In 3 of the remaining 4 winter cases and the 1 summer case a high from Iceland or Greenland drifted southward.

In these more persistent cases of type II the Highs in the summer six months are nearly always centered to the W. or N.W. of the British Isles or over the British Isles; relatively insignificant changes of pressure are sufficient to change the position of the centre within the area in question. In the winter the highs in about half the cases are in the same position; in other cases they may be centred over the Channel or to S.W. of Ireland, or farther W. on the Atlantic.

In type III, when persistent for 10 days or more, it is very often the case that the highs lie over Greenland or Scandinavia or drift eastward from Greenland, but there seems to be a liability to fairly rapid displacement.

In the matter of upper air temperature, since the highs in type II were in most cases centered to westward of the British Isles, the only temperature observations available are those made in the British Isles, i.e. in general in the northerly current. In summer in the cases of duration 10 days or longer the air is stable in the vertical and on the whole rather cold for anticyclones. Roughly the same - in rather more marked degree is true in winter. But if the temperatures in the southerly air currents on the west sides of the anticylones were available, the results would almost certainly show warmth as well as stability.

As regards individual cases some have been cold and some warm (on the east sides) without any apparent relationship between temperature and duration. The two highs which in winter drifted southward from Iceland were distinctly cold, especially in the lower layers; one which came from Greenland was cold in the lower layers and warm at higher levels. On the other hand some of the highs which have appeared to spread northward from the Azores were cold on their east sides at all levels up to the 450 mb. level, the apparent drift of the High northward from the Azores in these cases being inter-connected with the spread of a strong cold current from the north over the British Isles.

Related to the question of persistence of high pressure systems is the question of possible recurrence at more or less definite intervals. Though the normal circulation of the atmosphere of the Northern Hemisphere is much less regular than that of the Southerm Hemisphere the mean movement of the Northern air cap at the 4 km. level can be regarded approximately as made up of a solid rotation once in 19 days in winter and once in 38 days in summer, with certain superposed systems.* The chief of the superposed systems are, a belt of high pressure between latitudes 50° and 65° extending from the British Isles across Europe and Siberia; highs over N.W. America and over Algeria; lows to the south of Greenland and over the Eastern Pacific. At 6 and 8 km. in summer the mean rotation times are 33 and 23 days respectively.

In view of this result it seemed worth examining the question whether occasions of Type II or Type III that is, the main high pressure system in latitude 50° to 60° or northward of 60°, rather than in the usual latitude of the Azores, show any tendency to recur after periods of the order mentioned above. Clearly if there were something in the atmosphere (most likely the high atmosphere) which favoured the occurrence of higher pressure or lower pressure than usual and if that something persisted long enough, some recurrence tendency with a period of this order is to be expected. On the other hand the degree of persistence commonly shown by pressure systems other than the subtropical high and the infrequency of any long continued regular west to east travel in the northern hemisphere make it less likely that recurrences would readily be discovered at intervals equal to the rotation period of the circumpolar vortex. The conclusion reached so far is that the weather types as classified yield no evidence sufficient to confirm a tendency for displacements in latitude of the main high pressure system to recur with any regularity at definite intervals, but on the other hand they are not inconsistent with the possibility that peculiarities of the upper atmospheric circulation do at times persist long enough to recur at intervals of the order of the circumpolar rotation periods computed by Lee.

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* A.W.Lee. Q.J.50, 1924, pp.69-74.

The question of recurrence was examined also by means of other data tabulated in the course of the investigation - namely daily differences between maximum pressures in the Azores region and minimum pressures in the Icelandic region. These differences afford a measure of the total W-E flow of air towards Western Europe. Occasions during the winter half-year where this flow exceeded a certain level (about 25% over the mean winter level) in the mean for three successive days were marked and the intervals to the next important maximum were noted. From about 20 to 30 days after a maximum the recurrences were about twice as frequent as for other intervals, but this increase of frequency is not necessarily significant and in any case is not high enough to be taken into account in forecasting.

Whilst it is felt that a detailed examination of the structure and history of Highs is needed, there is a difficulty about Highs which does not apply to Depressions. Depressions, at least in winter, travel on fairly definite courses; they may cross the British Isles and it is possible from the records of a few continuously recording ground stations and with a limited number of upper air ascents to make fairly complete cross sections of the depressions. But anticyclones, especially persistent ones and excluding wedges of high, cover large areas - often ocean areas -, change only slowly and seldom follow definite paths. In none of the persistent cases selected above was it possible to get representative data for all parts of the anticyclone or to prepare any sort of cross section.

The following case illustrates another kind of difficulty. The most persistent anticyclone in lat. 50°-60°N. was that of April 1938, with absolute drought in the southern half of

England. This followed a March which was unprecedently warm, exceptionally dry and very sunny in England and was followed in turn by a May which was unseasonably cool and mainly dull.

Brooks (Met.Mag., June 1938) pointed out the probable explanation. Throughout January, February and March 1938 the pressure deviations from normal over Europe, Siberia and the Arctic were such that abnormally persistent southerly winds prevailed on the N. coast of Europe and western Asia. This southerly current of air drove the ice northward with the result that high temperature and low pressure prevailed over the Arctic in February, March and April. High pressure and drought over the British Isles during these months was the natural corollary. It would be interesting to apply the method of pressure waves to the first six months of 1938 but resources have not been available.

This case shows how wide a range of factors enters and how exhaustive information about weather types and the probable consequences of large deviations from the normal circulation is needed for an effective attack on the problem of long range forecasting.

WEATHER TYPES.

TYPE I: Anticyclone in Azores region (and usually also across to Bermuda).

- IA: without other important high pressure over Europe or N.Atlantic.
- IB: with high pressure also in the latitude of Greenland.
- IC: with extension towards Iceland.
- ID: with extension towards Russia (including extension over British Isles).

TYPE II: The main high pressure system in the latitude of the British Isles (50°-60°N), whether centered over the British Isles or W.Europe, or adjacent Atlantic region.

TYPE III: The main high pressure system north of latitude 60°.

TYPE IV. Isolated European anticylone centered south of 60°N., i.e. southerly or south-easterly type for the British Isles.

TYPE V. A very large anticyclonic area - a type occurring seldom except in mid-winter.

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METEOROLOGICAL RESEARCH COMMITTEE 10th March 1942

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MEMORANDUM.

Visibility and the Optical Properties of the Atmosphere.

Origin of Visibility Observations.

By G.C.Simpson.

The measurement of Visibility was introduced into meteorology in order to give practical information required by aviation. The aviator in planning his flights wishes to know not only the cloud and wind he is likely to meet; but also how far he can see, and he demanded that the weather reports should include this information. In order to meet this demand the meteorologist had no alternative at that time, but to observe how far he could see. This was simplified by the fact that all the easily distinguishable features of a landscape appear and disappear more or less at the same distance, and he took as his definition of visibility the greatest distance at which the features of the landscape could be distinguished as such.

For practical reasons, both of observing and reporting (by code figures), the meteorologist chose a number of distances and limited himself to observing and reporting whether features of the landscape at these distances could or could not be distinguished. The "features" which are to be distinguished have to be such that they are clearly visible in a clear atmosphere and under normal daylight illumination. Whether an object can be distinguished or not is a matter of whether its outlines can be seen sufficiently clearly against the background for one to be sure of what one is seeing; thus if the object is a tree the observer must be able to say that it is a tree and not a house.

The chosen distances were those of greatest practical importance; if one can see about 20 miles it does not matter much whether one can see 19 or 21 miles; but if one can see only 2 miles it is not sufficient to be told one can see 1 mile or 3 miles. Thus the nearer objects should be closer together than distant ones. It was also necessary that the distances should be nice round numbers and after international discussion the following distances, each of which is roughly twice the previous distance, were chosen:

Distance of visibility objects.	Code.
0	0
50 m	1
200 m	2
500 m	3
1 Km	4
2 "	5
4 "	6
10 "	7
20 "	8
50 "	9

So far everything was empirical and practical and there was no reference to the optical properties of the atmosphere.

Visibility in daylight.

There have been a large number of papers published dealing with the physics of visibility in daylight. A bibliography of these fairly complete up to 1939, will be found in the second edition of Middleton's book "Visibility in Meteorology". This book also contains a satisfactory summary of all the more important work on the subject and should be consulted for details of the work discussed in this memorandum.

The theory of daylight visibility worked out by Koschmieder is the one which is now generally accepted and is the only one which can be dealt with here.

Koschmieder's theory.

Koschmieder deals with the simplest form of visibility object, namely a perfectly black screen seen against the sky on the horizon.

The theory is very simple:

The cone of atmosphere extending from the eye of the observer to the contour of the screen is traversed by light from the sun, the sky and the ground. This light is scattered by the air molecules (Rayleigh scattering) and by the suspended matter. Some of the scattered light reaches the eye of the observer and appears to come from the screen so giving it an apparent surface brightness. This brightness depends on the amount of scattering between the eye and the screen and therefore on the distance of the screen; thus the surface brightness increases as the screen recedes from the observer. This surface brightness must obviously be less than the brightness of the aky adjacent to the screen, for the latter is made up of the light scattered by a cone practically similar to that in front of the screen, together with the light scattered in the extension of the cone to the limit of the atmosphere in the direction of the horizon.

Let H_1 and H be the apparent surface brightness of the sky and screen respectively, then as the screen recedes H_1 will remain constant and H will increase; the contrast between them is defined as $\frac{H_1 - H}{H}$ and it is a physiological law that the eye ceases to recognise this contrast when it falls below a small value called the threshold of contrast. At this point the screen becomes indistinguishable from the sky and the visibility range is reached. It was Koschmieder's problem to relate this range with the physical characteristics of the atmosphere.

The factors and assumptions used by Koschmieder.

Absorption and scattering. When a narrow pencil of light traverses the atmosphere its intensity decreases through scattering and absorption. Scattering is produced by material particles which in themselves are transparent but have a refractive index greater than that of free space. In the atmosphere these are the molecules of the air, the suspended water drops and a certain proportion of the nuclei. Nuclei may be divided into two classes (a) hygroscopic nuclei and (b) dust. The former are at all temperatures in the form of solutions in water and therefore are transparent droplets which scatter the light without absorbing it.

Absorption is confined entirely to the non-hygroscopic nuclei. These are usually particles of soot and mineral matter, both of which absorb all or the greater part of the light which falls upon them. There can be no doubt that in the atmosphere, except under the most highly polluted conditions such as occur in the neighbourhood of industrial regions or forest fires, the number of the non-hygroscopic nuclei is insignificant compared with the number of hygroscopic nuclei. Hence in unpolluted air the attenuation of a beam of light is entirely due to scattering and the absorption may be neglected.

This makes the solution of Koschmieder's problem relatively easy, for the production of scattered light is equal to the elimination of light from the direct ray. In other words the attenuation coefficient and the scatter coefficient are the same.

The attenuation or scatter coefficient, σ, is defined by the relationship

$$E = E_0 e^{-\sigma x}$$

in which E is the flux-density in a beam of light at a distance x which has the flux-density E_0 at $x = 0$.

It is very important to realise that the numerical value of σ depends on the unit of length used, and care must be taken that if distances are measured in feet the value of σ appropriate to feet, usually written $\sigma\, ft^{-1}$ is used.

Uniform atmosphere.

In order to make his equations tractable Koschmieder assumes that σ is constant throughout the cones of air he is investigating. This means that the physical condition of the air is uniform, at least along the line of sight to the horizon in the direction of the target. In the neighbourhood of towns this is often far from being the case and therefore great care should be exercised if Koschmieder's results are to be applied to observations made in industrial regions. It is also clear for similar reasons that visibility observations made on the ground cannot be applied directly to visibility from aeroplanes, even of objects on the ground, for σ varies very much in the vertical direction.

Uniform illuminations.

Koschmieder's analysis assumes that every element of the two cones of air receives the same illumination. This is only true in two cases (a) with a clear sky so that the sun shines everywhere and (b) with a completely overcast sky. As light also enters the cones from the ground a uniform surface is also necessary: thus there should be no large expanses of water between the observer and the object, and if there is snow it should not be in patches. With a uniform surface it does not matter whether the surface is of land or water, of light or dark earth or even of snow - uniformity is the criterion.

The threshold of contrast, ε.

The threshold of contrast varies from observer to observer and depends amongst other things on the general intensity of the light. As it is not practical to allow for these variations ε is assumed to be constant and to have a numerical value of 0.02 (as ε is a ratio it is independent of the units of length). Experience has shown that in ordinary daylight conditions and with observers of normal eyesight the errors due to assuming ε to be 0.02 do not exceed those due to other unavoidable errors of observation.

Koschmieder's results.

With these assumptions and limitations Koschmieder's analysis shows that the visible range V is given by

$$V = \frac{1}{\sigma} \log_e \frac{1}{\varepsilon}$$

If the kilometer is taken as the unit for V and σ, and ε is taken to be 0.02 this reduces to

$$V = \frac{3.91}{\sigma} \text{ km.}$$

Koschmieder also showed that this relationship is independent of the illumination so long as it is uniform in the sense described above. Also that it is independent of the position of the sun so that the range of visibility of a black screen against the sky at the horizon is the same when looking towards the sun as in the direction away from the sun, although in the former case the brightness of the sky is much greater than in the latter case. He confirmed these results by actual experiments with a large black screen seen against the sky.

Thus if all visibility objects were black screens rising above the horizon the attenuation (scatter) coefficient of the atmosphere could be determined from the visibility observations. Visibility objects are however seldom black screens seen against the horizon and we must now examine the errors due to this fact.

Application of Koschmieder's results to ordinary visibility objects.

A discussion of the visibility of objects other than a black screen seen against the sky will be found in Middleton's book and a very useful summary has been given by Wright in Appendix I to his paper on Atmospheric Opacity (H.L. Wright. Quar. Journ. Roy. Met. Soc. Vol.65, p.431, 1939). The results derived in these discussions only will be given here, the analysis and data on which they are based must be consulted in the works mentioned.

Objects seen against the sky but not black.

If the object is not black some light is added to the scattered light in the cone between it and the observer; thus the object would become invisible at a shorter distance than if it were black. So long as the albedo of the object is not more than .25 no error larger than 3% is caused if the sky is overcast; but if the sun is shining on the object the error may be much larger. It is therefore desirable to choose objects as black as possible: a white house on the top of a cliff would be quite unsuitable as a visibility object: but a weathered building or a mass of dark trees would be satisfactory except when brightly illuminated by the sun.

Objects seen against a terrestrial background.

We have to differentiate between two cases: (a) the object and the background both at the same distance from the observer and (b) the background at some considerable distance beyond the object. In the former case even a black object seen against a background having an albedo of .25 (i.e. more than the albedo of most terrestrial backgrounds) would produce an error of 50% with an overcast sky and more than 70% if the sun were behind the object. Thus a tree on the edge of a wood would not be a suitable object. In the second case the object is not seen against the real background; but against the scattered light between the object and the distant background. If the background is not closer to the object than one third of the distance of the object from the observer the error would not be more than 10% assuming that the object is reasonably dark. As the objects used as visibility marks usually stand well in front of their background only small errors are introduced.

Coloured objects.

As objects near the limit of visibility lose their colour, appearing grey, it does not matter what colour the object may be - it is not the colour, but whether the object appears light or dark which is the decisive factor.

Wright sums up his discussion of this aspect of the problem in the following words "We may conclude that although the observed visibility is always less than the standard visibility, the differences for the kind of objects used at most meteorological stations are only likely to be serious in the somewhat special circumstances of an object viewed in bright sunshine when the solar bearing is near a critical azimuth".

Thus from visibility observations werobtain an approximate value of the attenuation coefficient σ, the accuracy of which is limited more by the fact that only a few objects are used than by the departure of the individual objects from the standard of a black screen seen against the sky.

The values of σ corresponding to the standard distances of the visibility objects adopted in international meteorology are shown in Table I.

TABLE I.

1	2	3	4	5	6
Distance of Objects	Attenuation coefficient σ	Candle power of lights at the same distance as objects		Distance of lights of 100 c.p.	
		International	Calculated	International	Calculated
Km	Km^{-1}	c.p.	c.p.	Km.	Km.
0.05	78.2	0.13	0.05	0.10	**0.14**
0.2	19.5	0.9	0.7	0.33	**0.39**
0.5	7.8	3.5	4.3	0.74	**0.78**
1	3.91	10	17	1.3	**1.3**
2	1.95	35	70	2.3	**2.1**
4	0.98	100	184	4.0	**3.3**
10	0.39	420	1740	7.5	**5.9**
20	0.19	1250	6960	12.0	**8.0**
50	0.08	4500	43500	-	**11.0**

Visibility at night.

Owing to the absence of sunlight it is not possible to use the ordinary visibility objects at night and they have to be replaced by artificial lights. At first sight it would appear a simple matter to make the change, for one would simply report the distance at which lights could be seen; but unfortunately all lights do not disappear at approximately the same distance in a given state of the atmosphere as terrestrial objects do.

Two methods remain:

(a) to determine the distance at which a light of a given candle power can be seen

and (b) to determine the candle power of the light which can just be seen at a specified distance.

If the visibility were required only for practical purposes such as aerial navigation there would be no need to connect either the night or the day visibility with any property of the atmosphere. It would be sufficient to say that in daylight terrestrial objects are recognizable to such and such a distance and at night lamps of a specified candle power can be seen for so many miles. But for scientific and meteorological purposes it is desirable to connect up both day and night visibilities with the same optical properties of the atmosphere. In the international weather service this has been attempted by specifying the distance at which a 100 c.p. light can be seen at night when the atmosphere in daylight would have a given visibility, or alternatively the candle power of a light which can be seen for the same distance at night as terrestrial objects would be visible in daylight if there were no change in the atmosphere. The values of these quantities adopted by the International Meteorological Organization are entered in columns 5 and 3 of Table I.

We will adopt the same method and derive these values on the same principles as those used in discussing the daylight visibility.

General principles. In daylight a distant object becomes invisible when the apparent brightness of the object, mainly due to scattered light, bears a certain ratio to the apparent brightness of the background. At night a light becomes invisible for a totally different reason; namely because the flux-density at the eye due to the emitted light falls below a definite amount called the threshold of flux-density. This is a much more variable quantity than the ratio in daylight observations; it depends amongst other things on:

(a) the condition of the eye, whether light adapted or dark adapted
(b) the brightness of the background
(c) the presence of other lights in the field of view.

Fortzik has adopted 3.5×10^{-7} lux (3.5×10^{-11} lumens cm^{-2}) as the threshold of flux-density, while Middleton considers this to be on the high side and prefers 2×10^{-11} lumens cm^{-2}. For practical purposes it is better to choose a high rather than a low value and Fortzik's value will be used in this memorandum.

In discussing the daylight visibility it was assumed that the attenuation of the light was due to scatter and that there was little or no actual absorption of the energy. The same conditions will hold at night hence the attenuation coefficient with a given state of the atmosphere will be the same by day and night. This is the link which makes it possible to connect the observations of lights at night with the visibility range in daylight.

The distance at which a specified light can be seen. The illumination from a point source of light decreases with distance for two reasons:

(a) the spreading out of the light
and (b) scatter.

Thus we have $E = I_0 x^{-2} e^{-\sigma x}$

in which E = the flux-density
x = the distance of the light
I_0 = the intensity of the light in candles
σ = the attenuation coefficient.

Let S be the distance in kilometres between the observer and the light when it is only just visible, then $E = .35$ lumens km^{-2} and we have:

$$.35 = I_0 \varsigma^{-2} e^{-\sigma\varsigma}$$

$$\text{i.e. } \varsigma^2 e^{\sigma\varsigma} = \frac{I_0}{.35}$$

This equation can best be solved by writing it in the form:

$$2 \log \varsigma + \sigma\varsigma \log e = \log I_0 - \log \cdot 35$$

and solving graphically.

The values of ς for $I_0 = 100$ candles are entered in column 6 of Table I against each of the daylight visibility ranges.

The candle power of a light which can be seen at a specified distance. It is possible from the above equation to calculate the candle power of a light at a given distance which would just be visible with any given attenuation coefficient. This however has little interest except in the case of lights placed near to the visibility objects; when it would be useful to know what candle power these lights must have to disappear at night in the same state of the atmosphere as the objects disappear in daylight. In this case $= V$.

As before we have:

$$I_0 = \cdot 35\, \varsigma^2 e^{\sigma\varsigma}$$

$$\text{As } \varsigma = V, \ \sigma\varsigma = \sigma V = 3.91$$

$$\therefore I_0 = \cdot 35\, e^{3.91} \varsigma^2 = 17.5\, \varsigma^2$$

This means that the light at a distance of 1 km would be 17.5 c.p. and all other lights increased as the square of the distance. It is interesting to note that if such a series of lights were set up in a perfectly clear atmosphere they would all appear equally bright, giving at the observing post a flux-density of 17.5 lumens km^{-2}, i.e. 50 times the threshold flux-density; then as the air became hazy they would disappear one by one commencing at the most distant, and the last light visible would always be at the daylight range of visibility. The intensities of the lights for each of the international visibility ranges have been entered in column 4 of Table I. It will be noticed that for distances greater than 10 miles the required candle power is too large to be practicable.

An interesting variation of the result in the last paragraph is seen if we consider a parallel beam of light, such as a searchlight beam with no spread. In this case there is no attenuation due to spreading and we have

$$E = E_0 e^{-\sigma x}$$

$$\text{when } x = V, \quad \sigma x = \sigma V = 3.91$$

$$\text{thence } E = E_0 e^{-3.91}$$

$$\text{i.e. } \frac{E}{E_0} = .02$$

That is every parallel beam of light is reduced to 2% of its original intensity in a distance equivalent to the visual range in daylight.

In columns 3 and 5 of Table I there are entered the values of the candle powers adopted by the International Meteorological Organization corresponding to those in columns 4 and 6 respectively, the latter being the values derived in this memorandum. Middleton (Visibility in Meteorology pp.92-95) throws doubt on these international values and suggests that they should be calculated using the modern formulae. This has now been done, and it will be seen that there are large discrepancies between the corresponding values.

Although the theory used in this memorandum for calculating the visibility of lights at night is now generally accepted; there is practically no experimental confirmation of the results. So far as I know no experiments have been published connecting day and night visibilities and until that has been done there will bo no final solution of the problems involved.

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METEOROLOGICAL RESEARCH COMMITTEE.

6 March 1942

MRP 17

WIND SHEAR AND CLOUD. By D.Brunt.

The theory of the circulations set up in unstable fluids was first given by Rayleigh (Collected Papers, Vol.VI, p.441), and was later completed by Jeffreys (Proc.Roy.Soc.A. 118, 1928, p.195), and Gellew and Southwell (Proc.Roy.Soc. A.176, 1940, p.312). It was shown that a layer of fluid could remain in equilibrium, even in a state of nominal statical instability, with the denser fluid above, provided that

$$\frac{\rho_0 - \rho_1}{\rho} < \frac{AK\nu}{gh^3} \qquad \text{.................. (1)}$$

where ρ_0 is the density at the top, ρ_1 the density at the bottom of the layer of depth h, K is the coefficient of molecular diffusivity, and ν the kinematic coefficient of viscosity. A is a constant whose value depends on the boundary conditions, being equal to 1709 when the liquid is bounded by solid surfaces above and below.

For a compressible fluid such as air the inequality takes the form

JUL 4 1944

$$\frac{T_1 - T_0}{T} < \frac{1709}{gh^3} \qquad \text{.................. (2)}$$

where T, T_0 and T_1 are the mean temperature, the temperature at the top, and at the bottom of the layer respectively. More accurately, the inequality (2) should be written

$$-\frac{\partial T}{\partial z} - \Gamma < \frac{1709 K\nu T}{gh^4} \qquad \text{.................. (3)}$$

where Γ is the dry adiabatic lapse-rate, but with shallow layers of air (2) is a close approximation, Γ being negligible by comparison with the critical values of $\frac{\partial T}{\partial z}$

Experiments carried out in the Meteorological Department at the Imperial College of Science and Technology have verified the limiting inequality for air, in which the motion is made visible by tobacco smoke, and for CO_2, in which the motion was made visible by a slight fog of Ammonium Chloride.

The chamber in which the motion was studied has as its base a metal plate which can be heated electrically from below, the top surface being the plane base of a vessel containing water, through which the motion of the air can be seen and photographed. The temperature was measured by means of platinum resistance elements, placed close to the upper and lower boundaries of the chamber. A dust-free layer of air forms in immediate contact with the heated plate, and since the air from this layer circulates around the boundaries of the convection cells, it is readily possible to observe the form of the convection cells.

It was found that when the temperature difference $T_1 - T_0$ exceeded the limiting values deduced from inequality (2), air initially at rest produced a series of convection cells, of a polygonal form. In each cell the motion was downward in the centre, outward at the base, upward at the margin. and inward at the top. In an incompressible fluid, the direction of the circulation was the reverse of that in air. The diameter of the cells was from 2.5 to 3 times the depth of the fluid.

Until the temperature difference between top and bottom of the chamber reached the limiting value no motion could be seen with layers of depth of 7 mm. or more. Thus, when the depth of the chamber was 1 cm., no motion of the air could be observed until the temperature difference reached 11.4°C. The motion which was then observed in a series of polygonal cells, as described above.

When the glass top of the chamber was moved steadily across the chamber, giving a shear within the chamber, the character of the motion was changed, the precise form depending on the rate of shear as well as on the temperature difference between top and bottom of the chamber. The following effects were noted:-

(1) Unless cellular motion was present the top plate was moved, no motion was produced by the shear. Thus shear alone produces no vertical circulation.

(2) With large shear, the columns of convection were sheared into vertical planes, and the chamber was filled with longitudinal rolls, i.e., rolls in the direction of the shear, with alternate rolls rotating in opposite directions. This occurred with all differences of temperature exceeding the critical values. The width of a double roll was equal to the diameter of a single cell in the unsheared state.

(3) With moderate shear and a large difference of temperature, the polygonal cells were distorted by the shear, and tended to be arranged in lines along the direction of the shear.

(4) Moderate shear, and a difference of temperature not greatly exceeding the critical value, gave transverse rolls, with alternate rolls rotating in opposite directions.

(5) A small rate of shear, and a difference of temperature just exceeding the critical value, gave longitudinal rolls. As the difference of temperature was increased, the chamber became filled in part with longitudinal rolls, and in part with transverse rolls, with an overlapping region of rectangular cells.

When the depth of the chamber was reduced to 6 mm. or less, it was found impossible to form polygonal cells by heating the base. The reason for this is readily seen by a consideration of the form of the inequality (2) above. At first sight, the difference of temperature appears to be inversely proportional to h^3, but the factor $\kappa\nu T$ becomes important as soon as the difference of temperature becomes large. As the base is heated, the mean temperature and the product $\kappa\nu$ both increase, and the product $\kappa\nu T$ increases more rapidly than the difference of temperature between top and bottom of the chamber. in shallow chambers it is therefore not possible to satisfy inequality (2) by heating the base.

It was found possible to do so, however, by inverting the chamber, and cooling the top by means of liquid air, and polygonal cells were then readily formed.

When efforts were first made to form polygonal cells in shallow chambers, another type of motion was observed, not similar to that in the polygonal cell, but consisting of columns of convection, surrounded by sluggish descent. The effect of shearing the top plate was then to fill the chamber with elongated spindle-shaped forms, which never approximated to the long rolls obtained by shearing a chamber arranged to give polygonal cells. How far such results can be applied to clouds in the atmosphere is not at present clear, though shallow cloud layers sometimes resemble the patterns of motion obtained in such shallow chambers.

Application to interpretation of clouds.

(a) Statical Experiments: i.e. with no Shear.

If convection in the atmosphere is similar to the motion in polygonal cells observed in the laboratory, we should expect to find the centres of convection currents spaced at distances of from 2½ to 3 times the height to which convection extends. This is confirmed by observation, so far as the order of magnitude is concerned. The descending current surrounding the ascending column covers about four times as great an area as the ascending current, and the ascending velocity should be about 4 times the descending velocity. The most expert English Sailplane pilot (Mr.Philip Wells) gave me, as an estimate of reasonable averages of ascending currents, and of the surrounding descending currents, 9 ft/sec. and 2 ft/sec., which are in the right ratio. He also measured on one occasion the height to which convection extended, as shown by a very thin cloud which capped the vertical current, and also the horizontal extent of the ascending current. The ratio of the latter to the former was 0.56, which again agrees reasonably well with the pattern of the polygonal cell.

In the absence of a variation of wind with height, convection cumuli should be spaced out rather like the holes in the top of a pepper pot, the mean distance of adjacent cloud being from 2½ to 3 times the height of the top of the cloud.

(b) Experiments with Shear.

(1) When there is a rapid increase of wind with height, the clouds form in long rolls, with ascending motion over the whole of the cloud base. In all cases in which the present writer has been able to check the direction of motion, the roll clouds have been observed to move along the direction of their own length. This is a matter which could usefully be checked by observers at outstations. There is no theoretical necessity for such a rule.

Mr.R.M.Poulter once sent me an observation of irregularly spaced cumuli made one morning, at a time when the pilot balloon showed wind not changing with height; later in the day, when a pilot balloon showed a fairly rapid variation of wind with height, the clouds were cumuli in long parallel rolls.

(2) When there is only a moderate variation of wind with height, the usual type of cumulus cloud will be long rolls

transverse to the direction of shear. Slight local variations of wind will produce local variations in the direction of shear, and as a result such clouds are seldom straight rolls across the sky, but are usually very much distorted. Short roll clouds transverse to the wind (travelling sideways to their own length) are by no means uncommon.

(3) Clouds which are in the form of small globules arranged in roughly parallel lines indicate a moderately small shear of wind with height, but a fairly large degree of instability within the cloud layer.

(4) Clouds which show two systems of wavelike patterns at right angles to one another indicate a low rate of shear of wind and a small degree of instability. Such clouds, which are far more frequent than might be expected, are subject to rapid variations in form, as one or other of the sets of "waves" predominates and dies away again.

The direction of shear, or of the vectorial variation of wind with height, is parallel to the isotherms within the layer under consideration. Thus longitudinal rolls should lie along the direction of the isotherms, and transverse rolls should lie normal to the isobars. The main distinction between the two types is one of magnitude of shear, and the roll clouds should lie along the isotherms when these are crowded together, and normal to the isotherms when these are at greater distances apart.

The breaking-up of layer clouds.

When layer clouds break up and disappear, the first stage of the breaking up is the formation of small cells or rolls, usually at the edge of the cloud. The rolls will be longitudinal or transverse according to the magnitude of the rate of variation of wind with height.

Sheets of cloud may become unstable through radiative cooling at the top and heating at the base, or by subsidence of the air in which they are suspended. In the latter case, the clear air below the cloud is carried downward dry-adiabatically, while the air within the cloud is only warmed up at the saturated adiabatic rate. As a result, the base of the cloud becomes unstable. The first sign of subsidence may be the breaking up of sheet of cloud into small cloudlets or rolls, which should be recognised by the expert meteorologist as indicating subsidence, which will rapidly lead to the disappearance of the cloud. I have observed large sheets of cloud of stratus type covering 8/10 of the sky decrease to 1/10 in less than 20 minutes.

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Dr. Stone: Thanks

Par a, page 4: (1) Val

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figures 9 1/5 + 2 1/5

during summer.

(2) Value $2\frac{1}{2}$ to 3

large. Bumps a

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Arizona indicate

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of separate Cu clouds

is less ~~on many occa~~

in summer, I believ

Par b, page 4, sub-

base rule has little founda

Cloud is composed of seri

at all levels and over all

Orientation tendency seems correct

(over)

HEADQUARTERS ARMY AIR FORCES

ROUTING SLIP

DATE 18 Sept 44

TO: Mr. Stone

OFFICE OR SYMBOL AFRWX	BUILDING Pentagon
ATTENTION:	ROOM 5D278

FOR:

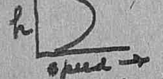
	COORDINATION		INVESTIGATION		NEC. ACTION
	INFORMATION		RECOMMENDATION		SIGNATURE
	FILE		CONFERENCE		

REMARKS: Except for the efforts to define the degree of instability, this article by Brunt is in close agreement with original experiments and findings of Walker in 1932.

In summary, Walker found that:

1) with no shear, polygonal cells (usually hexagonal) would be formed in unstable air.

2) with shear changing sign with increase in height as



, transverse rolls would be produced, with secondary rolls rotating in opposite sense from the lower primary rolls and alternately spaced above and between the latter.

3) With increased shear and of only one sign, longitudinal rolls were produced.

I believe this process is significant in the explanation of "cloud streets" associated with the tropical front

FROM:

NAME Col. McNeal		OFFICE OR SYMBOL AFACT-2
PHONE 4631	BUILDING Pentagon	ROOM 4C-1083

3-218, AF

---

of b.

", Koschmieder gives

$V_e = \frac{3.91}{\sigma_e}$

(over)

Dr. Stoal: Thanks much. Very interesting

Par a, page 4: (1) Values would seem to vary with air mass, season and latitude. Believe the figures $9\frac{1}{5}$ + $2\frac{1}{5}$ low for Texas during summer.

(2) Value $2\frac{1}{2}$ & 3 seems too large. Bumps at 12,000' over southern New Mexico and Arizona indicated a smaller value for that area at that altitude and during summer season. Proximity of separate Cu clouds in Texas is less ~~on many occasions~~ generally in summer, I believe.

Par b, page 4, sub-par (1): Believe the rule has little foundation in fact. Cloud is composed of series of vortices at all levels and over all areas.

Orientation tendency seems correct

(over)

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of b.

", Koschmieder gives

$$V_2 = \frac{3.91}{\sigma_2}$$

(over)

subpar (2): Transverse rolls exhibited best along discontinuity surfaces.

(3) – No comment

(4) An interesting phenomenon to watch. Saw such clouds appear and disappear 5 times in 20 min. one evening between 8 + 9 at Randolph.

D.B.B.

Ministry of Home Security

RC (G) 15.

MRP 18

CIVIL DEFENCE RESEARCH COMMITTEE.

SUB-COMMITTEE G.

(PHOTOMETRIC PROPERTIES OF THE ATMOSPHERE).

The Photometric Conversion Diagram.

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The Purpose of the Diagram.

The diagram has been prepared to represent graphically the relationship between the visible range, the attenuation coefficient and the transmission coefficient in the atmosphere. It can be used for any of the following purposes:-

(a) To convert any one of the visual range, the attenuation coefficient or the transmission coefficient into either of the other two: e.g. visual range into transmission coefficient; transmission coefficient into attenuation coefficient etc.

(b) To convert the unit of length from kilometre, mile or 1,000 yd. into either of the other two: e.g. attenuation coefficient expressed in miles into attenuation coefficient expressed in kilometres, etc.

(c) To combine (a) and (b) so that a given factor expressed in one unit may be converted into another factor expressed in a different unit: e.g. visual range in miles into transmission coefficient in kilometres.

The Construction of the Diagram.

Attenuation Coefficient σ.

σ is defined by $E = E_o e^{-\sigma x}$.

As this is true for all units of length we have

$a\sigma_a = b\sigma_b$ in which $a = x$ in units of a. ..... (1)
" " $b = x$ in units of b.

If V_a is the daylight visible range in units of "a", Koschmieder gives

$$V_a = \frac{3.91}{\sigma_a}$$

(over)

Substituting for σ_a from (1)

$$\sigma_b = \frac{3.91}{V_a} \cdot \frac{a}{b} \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots(2)$$

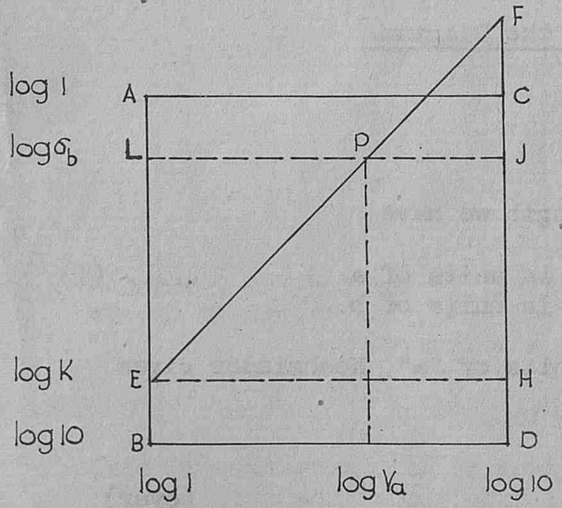
$$\sigma_b = \frac{K}{V_a} \quad \text{in which } K = 3.91\,\frac{a}{b} \qquad \ldots\ldots\ldots\ldots(3)$$

Values of K for different units.

V_a in Km .˙. a = Km.

a	b	$K = 3.91\,\frac{a}{b}$
Km.	Km.	3.91
Km.	Miles.	6.29
Km.	1,000 yd.	3.58

Geometrical construction on double log. paper.



Let AC represent log V_a from $V_a = 1$ to $V_a = 10$.

Let AB represent log σ_b from $\sigma_b = 1$ to $\sigma_b = 10$.

In AB take E so that AE = log K.

From E draw EF at angle 45°.

Let P be a point on EF, the co-ordinates of which are log V_a and log σ_b.

In the figure we have

$$AE = \log K.$$

$$FC = EB = 1 - \log K.$$

$$CJ = \log \sigma_b.$$

$$PJ = LJ - LP = 1 - \log V_a.$$

$$PJ = FJ = FC + CJ$$

$$PJ = FC + CJ.$$

$$1 - \log V_a = 1 - \log K + \log \sigma_b.$$

$$\log \sigma_b = \log K - \log V_a.$$

$$\sigma_b = \frac{K}{V_a} = \frac{3.91}{V_a} \cdot \frac{a}{b}$$ which is the required relationship.

Transmission coefficient, t.

$$t_b = e^{-\sigma_b}$$

$$\log e \; t_b = -\sigma_b$$

$$\log e \; \frac{1}{t_b} = \sigma_b$$

$$2.30 \log \frac{1}{t_b} = \sigma_b = \frac{K}{V_a} \quad \text{from (3)}$$

$$\therefore \log \frac{1}{t_b} = \frac{K}{2.30\, V_a} \qquad \ldots\ldots\ldots\ldots (4)$$

By giving t_b a series of values and choosing the appropriate values of K from the table, V_a can be calculated from (4). The values of t_b are then plotted against V_a and connected by a smooth curve.

The Use of the Diagram.

The daylight visible range, V, is given in kilometres by the scale at the bottom and in miles by the scale at the top of the diagram.

The attenuation coefficient, σ, is given by the scale on the right.

The transmission coefficient, t, is given by the scale on the left.

The optical state of the atmosphere is characterised by a vertical line on the diagram. Where this line meets the scale at the bottom of the diagram we have the visible range in kilometres and the corresponding visible range in miles where it meets the scale at the top.

The same vertical line cuts the three parallel straight lines (the σ lines) in three points; the readings on the right-hand scale against each of these points give the values of the attenuation coefficient, σ, in the units entered against the line.

Similarly, the same vertical line cuts the three curved lines (the t curves) in three other points. The reading on the left-hand scale against each of these three points gives the value of the transmission coefficient, t, in the unit entered on the curve.

Thus for the optical state of the atmosphere represented by any vertical line of the diagram we can read off at once:

V km^{-1}, V mile^{-1}, σ 1,000 yd^{-1}, σ km^{-1}, σ mile^{-1}, t 1,000 yd^{-1}, t km^{-1} and t mile^{-1}.

It is clear that conversion from one factor to another or from one unit to another can be effected at once by drawing the vertical line through the given value on its appropriate curve. It is then only necessary to read the scale value against the point where this line cuts the line or curve for the required factor and unit.

A line has been added to the diagram to represent nebules per kilometre (E.Gold Quar.Jour.Roy.Met.Soc.Vol.65, pp.139-164,1939). The relationship between nebules km^{-1} and $\sigma\,\text{km}^{-1}$ is

$$n = 14.5\,\sigma$$

The value of n for any point on the nebule line is to be read on the right-hand scale.

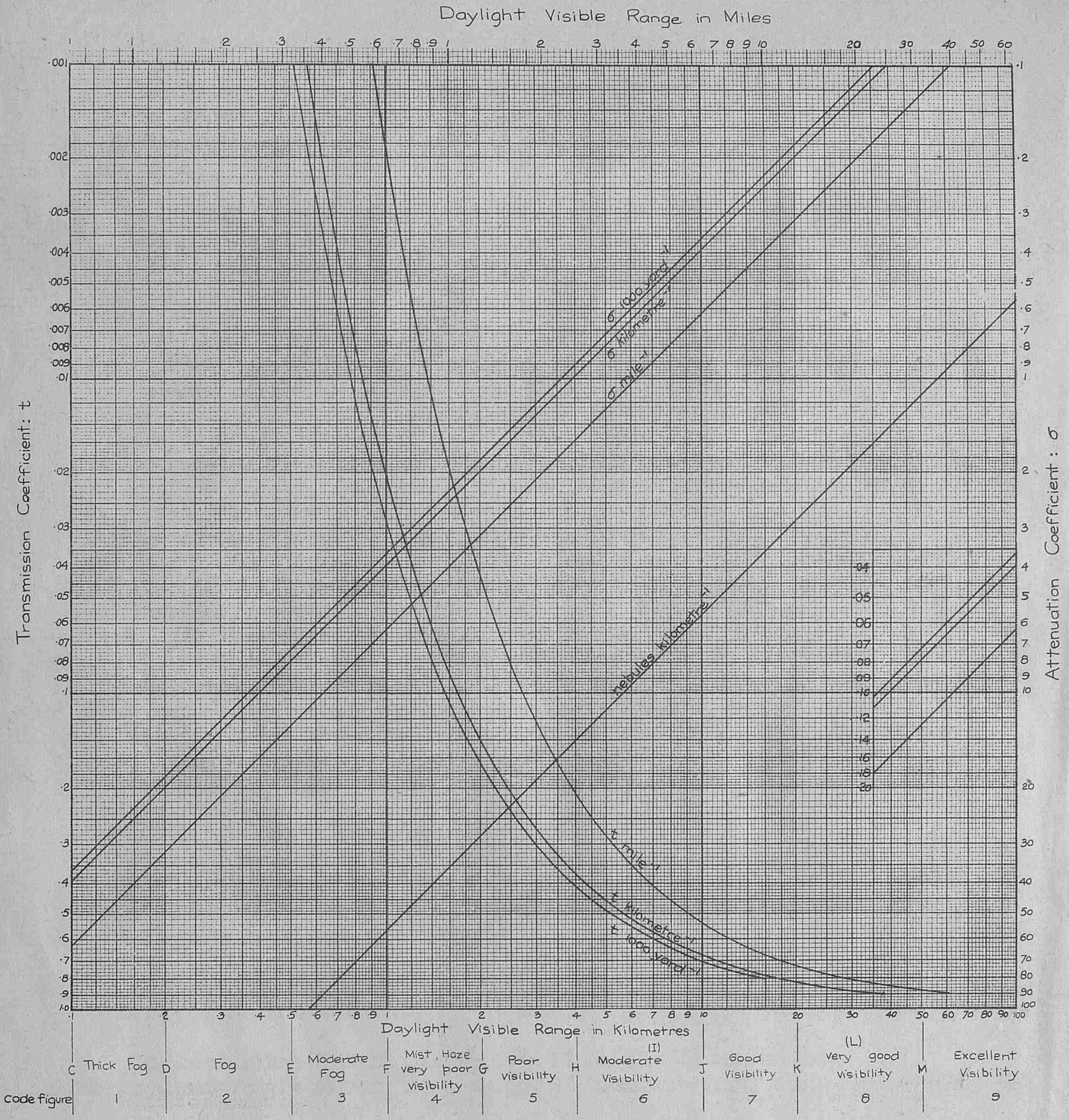
(A nebule is defined as a unit of obscuring power of such a magnitude that 100 nebules reduce the flux-density in a parallel beam to 1/1000th of its original value).

Kew Observatory. 21.3.42.

G.C. Simpson.

2739(27.3.42)

RC (G) 15.


Daylight Visible Range in Miles
·1 ·2 ·3 ·4 ·5 ·6 ·7 ·8 ·9 1 2 3 4 5 6 7 8 9 10 20 30 40 50 60
Transmission Coefficient: t
·001 ·002 ·003 ·004 ·005 ·006 ·007 ·008 ·009 ·01 ·02 ·03 ·04 ·05 ·06 ·07 ·08 ·09 ·1 ·2 ·3 ·4 ·5 ·6 ·7 ·8 ·9 1·0
Attenuation Coefficient: σ
·1 ·2 ·3 ·4 ·5 ·6 ·7 ·8 ·9 1 2 3 4 5 6 7 8 9 10 20 30 40 50 60 70 80 90 100
·04 ·05 ·06 ·07 ·08 ·09 ·10 ·12 ·14 ·16 ·18 ·20
σ 1000 yard⁻¹
σ kilometre⁻¹
σ mile⁻¹
nebules kilometre⁻¹
t mile⁻¹
t kilometre⁻¹
t 1000 yard⁻¹
·1 ·2 ·3 ·4 ·5 ·6 ·7 ·8 ·9 1 2 3 4 5 6 7 8 9 10 20 30 40 50 60 70 80 90 100
Daylight Visible Range in Kilometres
C Thick Fog
D Fog
E Moderate Fog
F Mist, Haze very poor visibility
G Poor Visibility
H Moderate Visibility (I)
J Good Visibility
K very good Visibility (L)
M Excellent Visibility
code figure 1 2 3 4 5 6 7 8 9

Gt. Brit. Met. Office 9561

AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE.

~~M.R.P.21.~~
30 April 1942.

Quarterly Report on Work done for the Meteorological Office during January to March 1942 at the Science Laboratories of the Durham Colleges, Durham.

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By E.Glückauf.

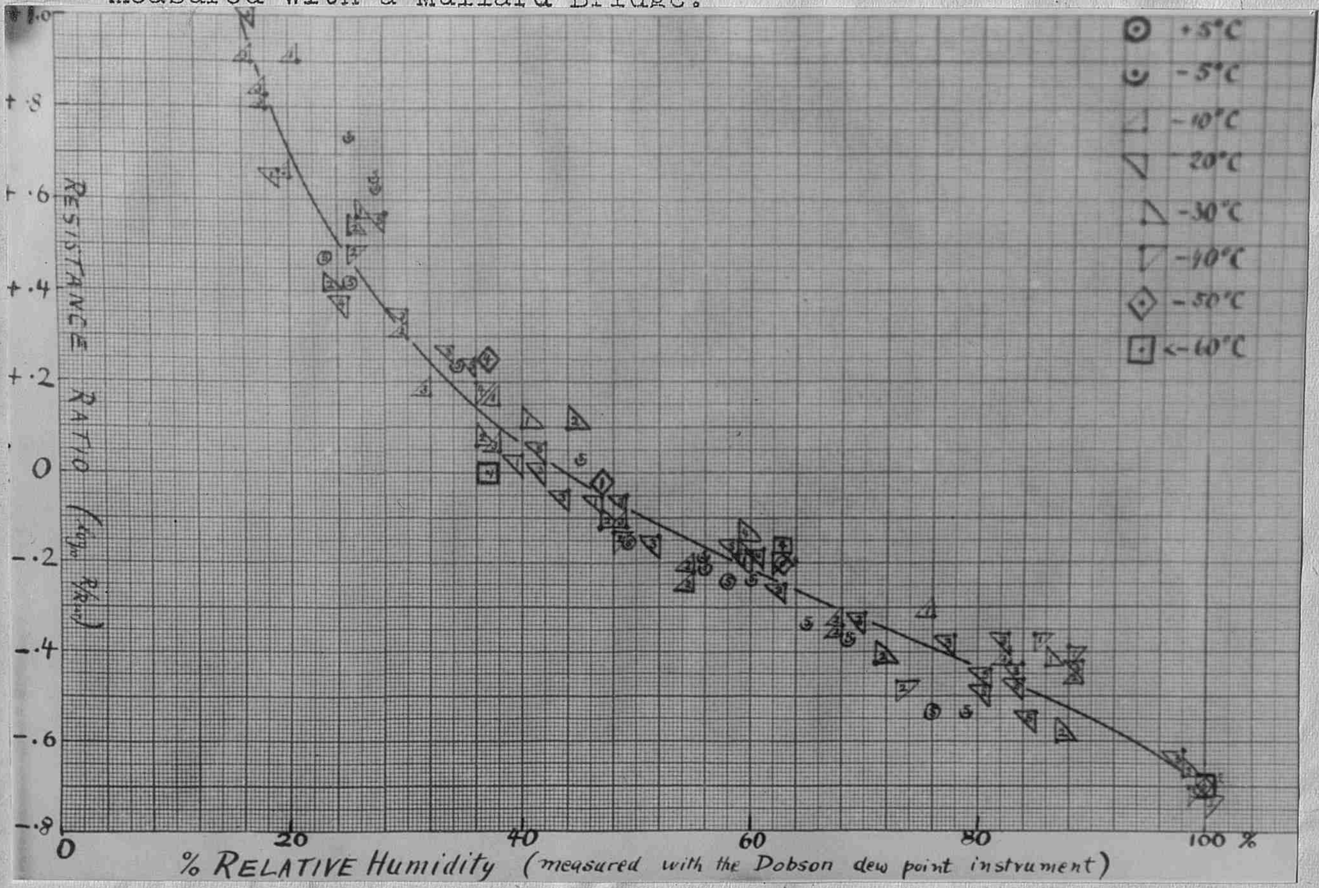
(A) Improvements on apparatus for testing hygrometers.

The Dobson dew point instrument was introduced for checking the relative humidity of the air passing through the testing chamber (see figure 1, last report), in which the desired relative humidity is produced by mixing in a definite ratio two air streams, one of which is perfectly dry, while the other is fully saturated. It was found that, at the high air velocities employed, full saturation was not obtained in the saturation spiral at temperatures above -25°C, probably on account of a "wet-bulb" effect produced by dry air passing over the frozen ice surfaces. Further more, the silica gel, after half an hour's use, was no longer sufficiently effective as a drying agent, so that the air after being cooled to -60°C. was no longer "perfectly dry". Within the range of -25 to -50°C, however, the relative humidities obtained by the method described in the last report proved to be perfectly correct. Once discovered, these faults in the testing apparatus were easily eliminated. The air coming from the silica gel absorber is now passed through a cooling spiral where the remaining traces of moisture are frozen out with solid CO_2 under a pressure of 130 atm. This trap also contains a plug of cotton wool, which prevents snow crystals from being carried along by the moving air. The air so dried was found to have a "hoar-frost" point of -89°C and thus could be considered as sufficiently dry for all experiments and calibrations down to -65°C.

The error resulting from insufficient saturation was allayed by introducing, for tests at higher temperatures, a washbottle filled with warm water between the stopcocks B and C (in figure 1, last report).

(B) Calibration of electric hygrometer against the Dobson dew-point instrument.

On account of earlier tests with the temperature-independent hygrometer the range of temperature constancy was believed not to extend above -20°C. After improvement of the testing apparatus, however, it was found that the range independent of temperature extends at least up to +50°C. To make quite sure that the independence of temperature is complete, series of calibrations were made whereby every relative humidity was determined by means of the Dobson dew-point instrument. The temperature-independent hygrometer used was of the type as shown in figure 9 (last report), and was dipped with a polyvinyal acetate solution (7% LiCl), diluted with another 20% of a solution of 50% alcohol-water. The actual calibration values are given in Figure 1. The ordinate shows the logarithm of the resistance ratio of the exposed part to the enclosed saturated part of the hygrometer, measured with a Mullard Bridge.



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3. 4.3.42.
4. 5.3.42.
5. 10.3.42.

Figure 1. Calibration curve of temperature-independent electric hygrometer. Relative humidities determined by means of the Dobson dew point instrument.

The diagram shows that there are no systematic variations of the resistance ratio over a range of temperatures from +5 to -70°C; nor were systematic variations due to ageing observed during a fortnight.

The lag constants of this element were the same as those observed with the American units (see table on page 11 of the last report). The unit which, therefore, appeared to be satisfactory, was then built into a screening case, for protection against rain droplets. This arrangement is shown in figure 2. The completed apparatus is now awaiting tests on aircraft.

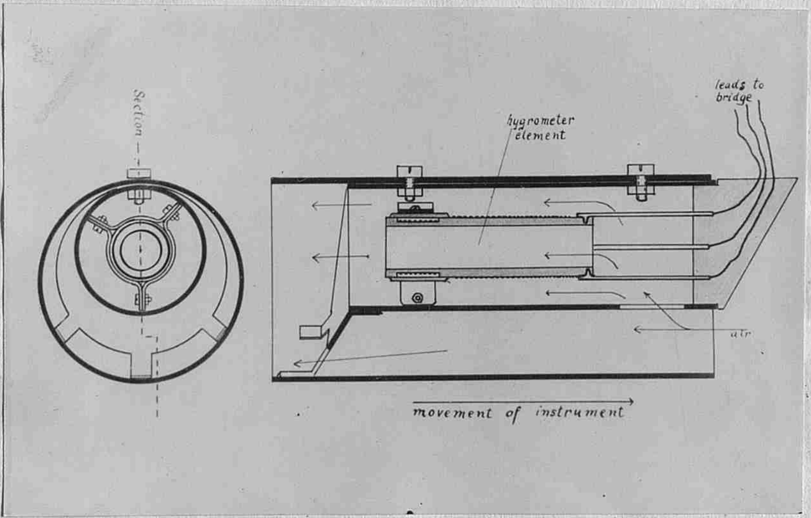

Figure 2. Arrangement of electric hygrometer in screening case for use on aircraft.

(C) Experiments concerning reduction of lag.

Considering that during the dipping process relatively large quantities of the solution must be trapped in the corners formed between the wires and the poly-styrene surface, platinised glass surfaces were tried instead of wire wound electrodes.

A first attempt to platinise a glass surface uniformly and to scratch by means of a steel needle a bifilar winding into the metallised surface, failed for mechanical reasons; moreover, the electrical resistance of the long and thin strips was too high.

More useful proved a method suggested by Mr.G.R.Martin to paint the electrodes directly on to the glass surface by means of the platinising solution. After some practice a number of useful units were obtained, both of cylindrical and of spherical form (see figure 3).

Experiments showed that the spherical form was superior in one important respect to the cylindrical one. If the latter is exposed to an air stream parallel to its axis, the lag in changing from high to low humidity is considerably bigger than vice versa. This effect is practically eliminated by using a half-sphere, which is fully opposed to the air stream, as shown in the lag curves of figure 4.

Cylindrical elements with their axis vertical to the air flow, where only the front half if platinised, would probably give the same advantage as the half-sphere, but the testing chamber is not big enough to mount cylindrical elements in this way; there is, however, no difficulty in calibrating them in a position parallel to the air flow, and to use them in a vertical position later.

Experiments also were made to see how much thinner the coating could be made without rendering the resistance of the unit inconveniently high. Dipping the element at a higher temperature (60°C) failed to give a reduction in the lag. Apparently the polyvinyl acetate solution becomes polymerised at higher temperatures, thereby compensating the decrease in viscosity.

However, dilution of the polyvinyl acetate solution with a 50% water-alcohol mixture resulted in coherent coatings which were visibly thinner than the usual ones, and which gave considerably smaller lag times, though producing somewhat higher resistances.

Two solutions were tested:

Solution No:	7A4	7A9
c.c. polyvinyl acetate solution	100	100
c.c. saturated LiCl solution	7	7
c.c. of 50% alcohol-water	400	900

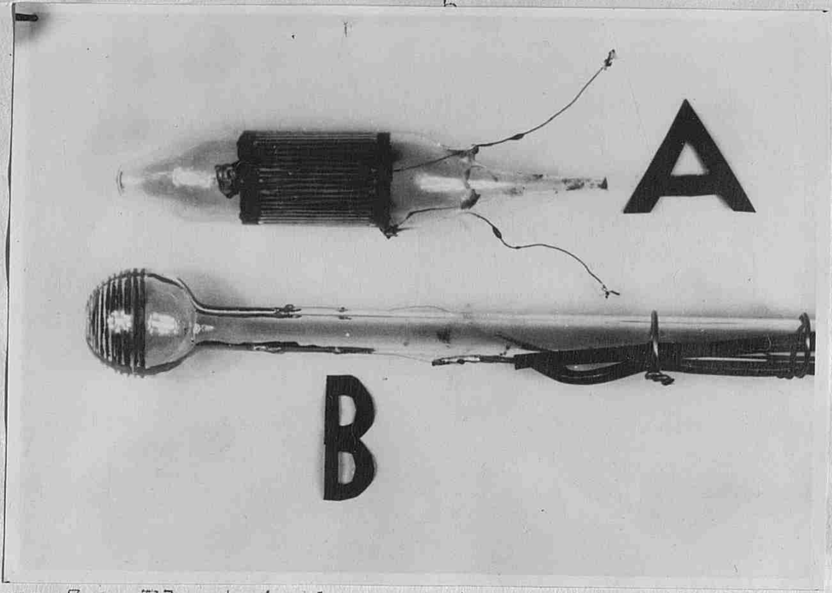

Figure 3. Electric hygrometer units of the "platinised glass" type (A) Cylindrical element (B) half-sphere element. (2/3 natural size).

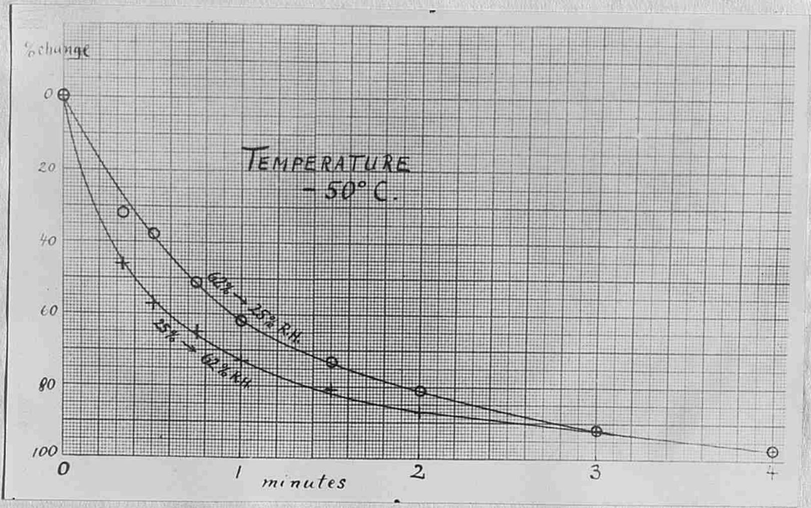

Figure 4. Lag curves of the half-sphere element shown in fig.3B, coated with solution 7A9. Maximum air velocity 40 miles/hour.

The results of these tests are given in the following table:

Temp. °C.	lag constant λ in minutes		
	usual sol.	7A4	7A9
-30	0.3	0.25	0.12
-50	3	2.5	0.8 x
-70	12 xx	-	9
Air speed	100 m/h	40 m/h	40 m/h

x see also figure 4.
xx probably too low (Caused by snow crystals)

In comparing these lag times due weight should be given to the difference in air speed (u), as the lag constant is approximately inversely proportional to $\sqrt{u}$.

The resistance of a half-shere element of 1" diameter at -70°C and 75% relative humidity (ice), coated with solution 7A9 is about 4 MΩ, which is still conveniently measurable (compared with 1 MΩ of the American unit under similar conditions).

While it is not possible to test this type of element at an air speed greater than that given in the above table, it may safely be expected that, at a uniform air speed of 100 miles/hour at -60°C. the lag time λ will be not more than about 1.5 minutes, i.e. about 1/5 of that of the American units.

(D) The Catalin hygrometer element.

A few tests were made with the Catalin hygrometer element, developed by Dr.Mann and British Thermostat Co., Ltd. The resistance of this element at saturation (below 0°C. with respect to ice) was measured. The values are given in figure 5. Even at saturation its resistance, when used with 50 ∿ A.C., is more than 10 MΩ at -30°C, and considerably higher at lower humidities. Moreover, the high capacity of the closely spaced windings does not permit measurements above 100 MΩ. It can, therefore, hardly be used as a hygrometer below - 20°C, and even then only at the highest relative humidities. The speed of response of the catalin element is certainly high; at -20^{C}C. it is too fast to be measured, but this is also the case for any element dipped into "solution 7A9" (see table above).

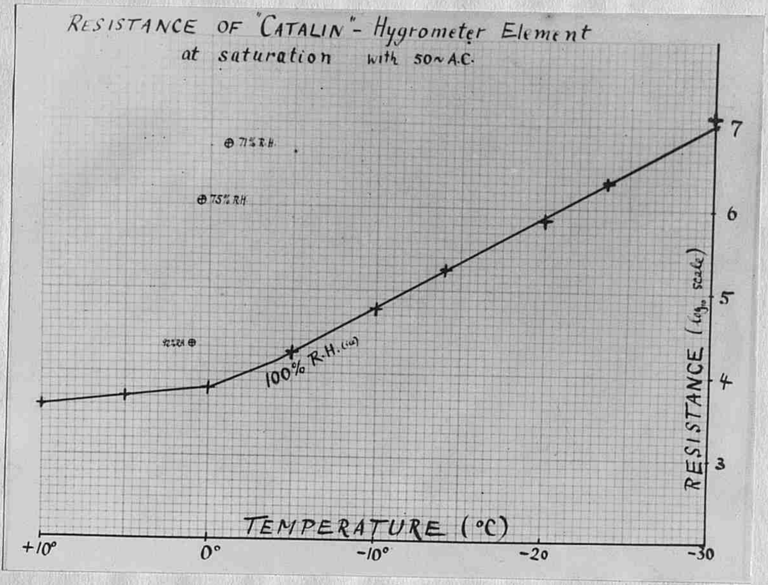

Figure. 5. Resistance of "Catalin" hygrometer element at saturation with 50 ~ A.C.

2 July 1942

METEOROLOGICAL RESEARCH COMMITTEE

JUL 4 1944

Determination of Atmospheric Transmission Coefficients of the Absolute Surface Brightness of the Sun as Functions of Wave Length.

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by H. H. Plaskett.

The work of Waldram is resulting in a determination at a median wave length (say λ 5500) of the atmospheric scattering coefficient as a function of height up to 30,000 feet, and of the atmospheric transmission coefficient from this height down to ground level. To complete this work on transmission so as to make possible a study of the radiative equilibrium of the whole atmosphere, particularly of that part above the tropopause, it is proposed that the transmission coefficients as a function of wave length be determined from outside space down to the level of the tropopause.

The investigation would be made by visual spectrophotometric methods (using a juxtaposed field to eliminate the effects of instrumentally scattered light), and the range of wave length covered would be λλ 4200-6700. A constant deviation spectroscope with telephoto lenses would provide a compact instrument, thus simplifying the design of a vibration free mounting, of sufficient dispersion and resolving power to open out widely regions of pure solar continuous spectrum free from absorption lines. As a secondary standard a small standardized coil filament lamp run against a voltmeter would be used, and this would in turn be calibrated on the ground against standard ribbon filament lamps (standardized at the N.P.L., at the Nela Labs., and in the Physics Department at Utrecht to give surface brightness in ergs. per cm.^{2} as a function of wave length). The instrument would be used in the wireless operator's seat in a Boston with the hatch open, and observations on any one flight would be limited to about an hour's duration while the sun's zenith distance was changing from 80° to 60°. If allowance were made for curvature (see Bemporad) this would provide an accurate determination of the atmospheric transmission coefficient down to the level reached by the aircraft, and the shortness of the duration would minimize any errors due to changing transmission with time, probably small at heights above

25,000 feet. Incidentally such observations would also yield the absolute surface brightness of the sun, the problem in which I personally am more particularly interested, with an accuracy far greater than can be reached by the usual method of determining the solar constant for total radiation, and then applying uncertain corrections for extrapolated solar spectrum beyond the limits of the bolometer spectrum.

The necessary parts of the spectrophotometric equipment, including the standard lamps, are already largely available at the University Observatory, and the mountings for these parts could be constructed there. What is sought from the Meteorological Committee is permission and authority to carry out the proposed investigation, that is to make the necessary eight to ten flights to a height of 30,000 feet in a Boston from Boscombe Down. Apart from the flying time, which would not be excessive, the greater part of the work would lie in the calibration of 'neutral' tint wedges, filters and of the secondary standard lamp. This would have to be carried out in the evenings after a flight had been made, and so would not interfere with the navigational research for which I am responsible at Boscombe Down.

University Observatory,
Oxford.
21 May 1942.

2. Letter from Professor Plaskett to Dr. Dobson, dated 21st May 1942.

"I enclose herewith an outline of the scheme I mentioned to you on Saturday. I fully realize that it is a piece of basic, that is long term research, and the only reason I put it forward is that under existing circumstances it could be carried out by me with very little interference to more urgent problems. As I point out in the outline the greater part of the work comes in calibrations after each flight is over, and since to do these calibrations I should have to have darkness and the office to myself this would necessarily mean that the work would have to be done after office hours, So that apart from flying time my navigation research would not be interfered with.

....

The advantages in the job being done by me are first that I have at the observatory most of the necessary equipment, secondly that I have some experience with photometric work and with observing in the open hatch of a Boston at heights from 25 to 30 thousand feet, and thirdly that for astronomical reasons I am perhaps far keener on the problem (and therefore willing to put up with more discomfort) than the average 'stooge' who might be assigned the job. Provided permission were given by the Committee to undertake the problem, I estimate that it would probably be six months before I could hope to have the apparatus assembled and tested. This would bring us well into the winter when the tropopause is at its lowest, so that on some of the flights I might hope to reach it at the ceiling of the Boston. On other occasions I would of course be below, but I should observe just the same. The resulting mean transmission coefficients would therefore refer to heights some 5000 feet below the tropopause.

From a meteorological point of view, if I might venture an opinion, I think such an investigation would be likely to be of value in supplementing and extending Waldram's results. From an astronomical point of view, and of course this is why I am keen on the problem, there is for the first time at least the opportunity of learning what the real energy distribution of the sun is."

3. Extract from letter from Professor Plaskett to Dr. Dobson dated 28th May 1942

"The Smithsonian determination of the solar constant gives the surface brightness of the sun in total radiation, averaged over the disc. The proposed investigation would give the surface brightness in momochromatic radiation, also averaged over the disc. Since the latter depends upon observations with a single instrument (in place of the pyrheliometer and spectro-bolometer with their independent errors), and is free from errors arising from uncertain allowance for water vapour absorption, as well as those arising from extrapolation at the ends of the spectrum, it is evident that a determination of monochromatic surface brightness should be more accurate than one made in total radiation. Further the short period of observation at the proposed considerable heights would greatly reduce the effects of secular changes in atmosphere transmission, already known from Linke's work to be affecting the Smithsonian results. The constancy (or variability!) of the surface brightness of the sun could therefore be established with relatively greater freedom from systematic errors in the proposed method. The accidental error in the proposed method would, however, be larger than that claimed by the Smithsonian people, and I would count myself lucky if it could be held at one per cent."

4. Extracts from letter from Professor Plaskett to Dr. Dobson dated 29th June 1942:

"Thank you for your letter of the 26th and the very excellent news which it contains. The proposed informal arrangement is just what I had hoped for, and seemsto me much the best way of getting the job well in hand." .....

"The first task for me is the collection of the necessary apparatus, and the construction of a suitable spectrophotometer. In my letter to you of 21 May I estimated that that would probably take six months. I am now writing around to the various firms for the bits of this apparatus which we have'nt got at Oxford. I am hoping that in most cases they will be able to deliver from stock, but where this is not possible it will probably be necessary to get some form of priority order, especially since the apparatus is being purchased by the Observatory, and not a Government Department. As soon as I learn what the situation is, and in which cases it will be necessary to have a priority order to secure delivery within say, eight weeks, I shall write and let you know. At its meeting on August 6th, the Committee could then possibly do something to expedite delivery on outstanding items."

U.S.

METEOROLOGICAL RESEARCH COMMITTEE — 27th July 1942

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Notes on Dew-point hygrometers designed at Oxford for use on Aeroplanes

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TECHNICAL REFERENCE FILES

by Dr. G.M.B. Dobson and P/O. Cwilong

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MRP 39

(1) Methods of cooling. For instruments which have to be used on single-seater aircraft at great heights it is of the utmost importance that they should involve the absolute minimum of action on the part of the Pilot, even if some accuracy is sacrificed since, if too difficult to use, no results at all will be obtained. Three methods of cooling have been used:-

(a) In the first method a jet of CO_2 from a cylinder is directed onto the under part of the test-plate. Cooling in controlled by opening and shutting the tap on the CO_2 cylinder. This is very simple but it is
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and the test-plate is cooled by thermal circulation, the cooling being regulated by an adjustable valve. An electric heater round the test-plate enables it to be warmed up at any required rate. In using the instrument the valve is set so that the test-plate slowly cools down and as soon as dew or ice is seen on the plate the heating is started and when the dew has evaporated the heating is stopped and the test-plate begins to cool down again.

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METEOROLOGICAL RESEARCH COMMITTEE 27th July 1942

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Notes on Dew-point hygrometers designed at Oxford for use on Aeroplanes

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(a) In the first method a jet of CO_2 from a cylinder is directed onto the under part of the test-plate. Cooling in controlled by opening and shutting the tap on the CO_2 cylinder. This is very simple but it is difficult to get very fine control of temperature owing to the tendency for the tap to block with solid CO_2 and then to come with a burst.

(b) In the second method (see A.R.C. report 5407 T.A.172) a jet of cold alcohol - cooled by contact with solid CO_2 in a vacuum flask - is forced against the under side of the test-plate, by air pressure supplied by squeezing a small rubber bulb. In this way it is possible to get a very accurate control of the temperature, but it is considered too difficult for use in single-seater aircraft. Hence a third method was tried.

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(c) A large vessel, well lagged, contains about 300 c.c. of solid CO_2 and alcohol and serves as a store of 'cold'. An inner vessel contains pure alcohol and is cooled by the 'cold store'. Pipes are led from this inner vessel to the test-plate which is at a lower level and is hollow and the test-plate is cooled by thermal circulation, the cooling being regulated by an adjustable valve. An electric heater round the test-plate enables it to be warmed up at any required rate. In using the instrument the valve is set so that the test-plate slowly cools down and as soon as dew or ice is seen on the plate the heating is started and when the dew has evaporated the heating is stopped and the test-plate begins to cool down again.

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Construction of the test-plate. At present this is made of copper with a solid gold sheet soldered to the top. The top is made fairly thick (about 3 m.m.) in order that the temperature conduction shall be good and all parts shall be at sensibly the same temperature. Three thermo-junctions are fixed in small holes which end just under the mirror surface of the test-plate. One junction is soldered in with Wood's metal but the other two have necessarily to be insulated. So far it has not been found possible to use any other metal than gold for the test-plate since other metals will not keep bright and allow the surface to be easily cleaned if dirt gets on them. The colour of gold is much better for seeing the deposit than that of silver or platinum etc.

Measurement of temperature. One requires to know accurately the difference between the temperature of the free outside air and that of the test-plate when dew just begins to be deposited. Three thermo-junctions are therefore fixed under the test-plate as described and the other three junctions are fixed in a 'Pitot' head outside the aeroplane, the Pitot head having small holes to allow some circulation of air - a method suggested by the N.P.L. Allowance must, of course, be made for heating by compression. The temperature should be read with an accuracy of at least 1°C by this method.

Detection of dew or frost. At the higher temperatures above say -50°C this presents no difficulty but at very low temperatures the amount of water vapour in saturated air is so small that special precautions must be taken to see that the most minute traces of deposit can be easily seen. If this is not done the test-plate may be cooled some degrees below the dew-point before a deposit is recognised. This difficulty is overcome by providing two inlets for the air entering the instrument: one is a large inlet and its purpose is to keep the whole of the air in the instrument constantly changed so that it may be an accurate sample of the outer air, the second inlet is a jet which directs a narrow stream of air onto the test-plate. When the test-plate is cooled to the dew-point a deposit forms in a narrow band across it and is easily seen. The jet has the further advantage that it brings fresh air constantly into intimate contact with the test-plate. The driest air that we have used had a Hoar frost-point of about -80°C. but no difficulty was found in getting a reliable temperature. (For such low temperatures in the laboratory liquid air was used instead of CO_2. In an aeroplane CO_2 should give sufficiently low temperature).

Tests of the accuracy of the instruments. Dry air was obtained by passing a stream of air from a compressor through a very large volume of dry silica gel. Various tests were made. (1) Cooling the test-plate moderately fast (say 2°C per minute) or very slowly indeed gave no difference. (2) Large changes in the velocity of the air passing through the instrument made no difference. (3) Large changes in the temperature of the air supplied to the instrument made no difference to the indicated dew-point. At the higher temperatures one can measure the temperature at which dew just begins to appear and that at which it just begins to evaporate and the two agree within a few tenths of a degree. (The temperature control with method (b) above is sufficiently good to keep the test plate at a constant temperature within a few tenths of a degree so that one can see the deposit begin to grow and begin to evaporate alternately). At very low temperatures when the deposit is ice the temperature at which evaporation is first noticed is some degrees higher than that when the deposit is first seen even if only a very minute amount of ice is allowed to deposit. The temperature of the test-plate when the deposit is first seen is much more constant than that when evaporation is first noticed and we think that it is very closely the true dew - (or hoar frost-) point. A test was made in which air of a hoar frost-point a little above -79°C was passed through a metal tube cooled in solid Co_2 and alcohol. This air was found to have a hoar frost-point very closely equal to the CO_2 temperature, if the temperature was measured when the first trace of deposit was seen.

A small trouble is that dust tends to be deposited on the test-plate by thermal precipitation and this makes it difficult to detect the first trace of dew after the instrument has been used for some time. Probably this trouble will be less when using the relatively clean air at a great height.

As has been mentioned, one thermo-junction is soldered to the test-plate while the other two are necessarily insulated electrically from it. In order to see whether the insulated junctions lag appreciably behind the true temperature of the test-plate the soldered junction and one insulated junction were connected in opposition so as to measure the difference of temperature between them.

Even when the test-plate was cooled rapidly the difference was not more than 0.5°C so that at the usual rates of cooling the error should not matter.

We have mentioned the difficulty of knowing whether the deposit is in the form of dew or ice. So far we have not got over this trouble. If the test plate is coated with zinc the deposit more easily forms as ice, but the zinc is not a good colour against which to see minute traces of deposit and it will not remain bright for long. In the laboratory one can always make sure if the deposit is water by cooling the test-plate further when a noticeable change takes place as the water freezes. This is probably too difficult an operation to do in the air since one must watch the deposit carefully to observe the change. Again in the laboratory one can make a small deposit of ice as above and allow it nearly to evaporate but start cooling before all the ice has gone. The remaining trace of ice acts as a nucleus on which fresh ice will form as soon as the test-plate reaches the hoar frost-point. Again this is probably too difficult an operation to do in the air.

Lead-in Tubing. On an aeroplane the air must be led to the instrument from outside through metal tubing. On a Spitfire nearly 15 ft. of tubing is necessary. In the laboratory it has been found necessary to clean the inside of such tubing carefully otherwise water vapour may be absorbed or given off. On an aeroplane where the temperature of the tubing may change by quite a large amount this will be very necessary. (Cleaning with chromic acid was found successful). Rubber tubing even if carefully dried for 24 hours continued to give off water vapour and rubber should not be used.

In use in the air a difficulty will almost certainly arise owing to water drops or ice particles entering the tubing if the aeroplane passes through even a trace of cloud. There are also conditions when hoar frost forms on the exposed parts of the aeroplane when flying through clear air: in such cases ice may also get into the tubing. Owing to the very slow evaporation at low temperatures ice entering the tubing at the start of the flight may invalidate the readings for the rest of the flight. Hence it would seem necessary to fit an airtight cap over the inlet which can be removed, say by Bowden control, when readings are being taken.

METEOROLOGICAL RESEARCH COMMITTEE 26th August, 1942

Investigations on Falling Drops carried out at the Chemical Defence Experimental Station, Porton.

by O. G. Sutton

Introduction

The behaviour of drops falling through the atmosphere has been investigated by the Chemical Defence Experimental Station, Porton in considerable detail over a number of years. It is proposed in the present memorandum to summarise very briefly the results of these investigations which although primarily designed for the study of methods of chemical warfare have considerable interest for meteorologists.

The principal lines of research have been concerned with

I. The determination of the terminal velocities of drops at heights up to about 20,000 feet.

II. The effects of atmospheric conditions on falling drops (evaporation and temperature changes).

The work has been shared between the Physics and Meteorological Sections of the Station. All the laboratory measurements have been carried through by the Physics Section which has also made some theoretical investigations. The Meteorological Section has been concerned with the investigations on the mathematical physics of the problems and has undertaken the bulk of the field work. What follows should therefore be regarded as the joint work of both Sections with the Physics Section as the prime mover.

I. The Terminal Velocities of Drops

1. Measurements

When a rigid body is released from rest in air it will fall with increasing speed until its velocity reaches a steady value called the "terminal velocity" i.e. when the weight of the falling body is exactly balanced by the air resistance. If drops of liquid were invariably spherical in shape the determination of the terminal velocity could be easily effected from examination of the well known curve which shows the drag of a sphere as a function of the Reynolds number.

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In practice it is known that if the diameter of a drop exceeds about 1 mm. it ceases to behave as a rigid sphere. There are thus three types of motion:-

(i) Motion of very minute drops (such as are found in mists) which obey Stokes' Law and whose velocity can be calculated exactly from the simple formula

$$v = \frac{g(\rho_1 - \rho)d^2}{18\mu}$$

(ρ = density of air, ρ_1 = density of drops, μ = viscosity of air, d = drop diameter).

(ii) Motion of drops whose diameters lie between 10^{-1} mm. and 1 mm. (e.g. fine rain). These have velocities outside the range of validity of Stokes' Law but otherwise behave as rigid spheres. No simple expression has been found to express the terminal velocity, but it is always possible to evaluate the terminal velocity from a knowledge of the drag co-efficient of the corresponding rigid sphere.

(iii) Motion of 'large' drops, whose diameter exceeds 1 mm. (rain). The terminal velocities of such drops may be as much as 50% lower than those of rigid spheres of equal volume, the deviation being primarily due to deformation from the spherical shape caused by the excess pressure forward giving an increased cross section and hence increased air resistance. It is also thought that at least two other factors complicate the situation, namely a possible slip effect brought about by circulation of the liquid (leading to a reduction of drag) and the increase of cross section due to the centrifugal forces set up by the internal circulation (leading to an increase of drag). An exact calculation of the terminal velocity of a large drop is therefore unlikely and recourse must be made to measurement.

Experimental determinations of the velocities of drops falling in air have been published by P. Lenard (Met. Zeit. 39, 249, 1904), W. Schmidt (Met. Zeit. 26, 183, 1909) and W. D. Flower (Proc. Phys. Soc. 40, 167, 1928). The last named worked at Porton.

The results of Lenard and Schmidt are shown with the Porton results in the attached graph. The Porton velocities are much higher for the larger drops. Flower's velocities are approximately correct, but they are not the terminal velocities since the height of fall (350 cm.) was inadequate. He designed a kind of ballistic pendulum to detect acceleration, but it was evidently insufficiently sensitive..

Further experiments were carried out at Porton using both Schmidt's method and the cinematograph method, but these cannot compare as regards accuracy with the photo-electric system at present in use. A further improvement in accuracy has been obtained by new methods of determining the size of drop which was formerly found as a function of the size of stain produced by a drop of dyed liquid on filter paper.

The latest type of apparatus for terminal velocity measurements is described in Porton Report No.2102. It was developed from an earlier type (Porton Report No.1883) and consists of a fall tube 23 cm. in diameter and 1100 cm. long. This is provided at the bottom with six pairs of windows through which narrow sharp edged beams of light pass. These are parallel and horizontal and fall upon vacuum photoelectric cells. The distance between adjacent beams is about 30 cm.

The output from the cells is fed through a condenser coupled amplifier to a double fibre Einthoven galvanometer. The movements of the fibres are recorded photographically. The passage of a drop through the beams of light cause one of the galvanometer fibres to execute six damped oscillations. The other fibre is connected to a 1000-cycle elinvar fork which is valve maintained and provides a time base. The velocity is found from the time intervals between the maxima on the photographic records for the first fibre and the distance between the light beams. The object in using six beams is to provide a delicate check on the absence of acceleration and to give a better mean velocity. The tube can be pumped down to about half an atmosphere to partially stimulate high altitude conditions; there is no temperature control.

The drop is formed from a jet (tip), and its size can be calculated accurately from the size of jet in the case of liquids fulfilling a certain relationship and released under certain conditions, from the experimental data of Harkins and Brown (J.Am.Chem.Soc. 51,499, 1919). In many cases when extreme accuracy is desired, these conditions are either doubtful, or not fulfilled (the error being up to 5% in the case of large drops), and the drop size has therefore been measured in recent experiments by using either a microburette in the case of small drops, or a photographic apparatus, for larger ones.

The order of accuracy aimed at in these experiments was 1%. This has been attained without difficulty.

The five velocities calculated from the successive intervals were constant within 1% except for large drops ($>$0.4 cm. diameter) of water at ½ atmosphere pressure when a total velocity increase rising to 5% in 200 cm. was noted.

The following tables contain all the experimental results for water. The air temperature throughout was between 17°C and 21°C.

(In the following ν = kinematic viscosity of air, ρ = density of drop, T = surface tension, g = acceleration due to gravity, all in C.G.S. units).

TABLE I

Velocities measured at 1 atmosphere corrected to 20°C, 750 mm.Hg.

Diameter of equivalent sphere, d.	Velocity of drop, v.
cm.	cm/sec.
0.069	278
0.077	318
0.079	314
0.229	697
0.338	827
0.407	876
0.442	899
0.493	902
0.542	918
0.595	919
0.624	909
0.637	914
0.667	917
0.725	916

The attached graph compares these figures with previous measurements by P. Lenard and W. Schmidt.

TABLE II

Experimental results over a range of pressures

Diameter cm.	$\log_{10} v$	Air density g/cm^3	Velocity cm/sec.
0.338	$\bar{1}.511$	0.000555	1080
	$\bar{1}.430$	669	1030
	$\bar{1}.348$	807	960
	$\bar{1}.283$	938	913
	$\bar{1}.212$	1104	857
	$\bar{1}.179$	1191	827
0.407	$\bar{1}.545$	0.000516	1150
	$\bar{1}.388$	741	1033
	$\bar{1}.314$	878	986
	$\bar{1}.263$	988	943
	$\bar{1}.212$	1112	901
	$\bar{1}.171$	1222	867
0.442	$\bar{1}.516$	0.000551	1160
	$\bar{1}.445$	650	1110
	$\bar{1}.354$	800	1025
	$\bar{1}.292$	924	990
	$\bar{1}.230$	1066	933
	$\bar{1}.182$	1189	900
0.493	$\bar{1}.484$	0.000593	1150
	$\bar{1}.364$	784	1065
	$\bar{1}.283$	944	995
	$\bar{1}.228$	1071	951
	$\bar{1}.198$	1148	921
	$\bar{1}.180$	1196	900
0.542	$\bar{1}.484$	0.000591	1175
	$\bar{1}.380$	751	1100
	$\bar{1}.314$	873	1040
	$\bar{1}.263$	982	994
	$\bar{1}.221$	1082	952
	$\bar{1}.184$	1180	921
0.595	$\bar{1}.535$	0.000525	1250
	$\bar{1}.435$	662	1160
	$\bar{1}.353$	799	1090
	$\bar{1}.274$	957	1010
	$\bar{1}.226$	1075	960
	$\bar{1}.181$	1185	921
0.624	$\bar{1}.536$	0.000527	1250
	$\bar{1}.421$	634	1180
	$\bar{1}.366$	778	1092
	$\bar{1}.298$	916	1026
	$\bar{1}.227$	1075	954
	$\bar{1}.189$	1171	917

Diameter cm.	$\log_{10} v$	Air density g/cm^3	Velocity cm/sec.
0.637	$\underline{1}.491$	0.000592	1190
	$\underline{1}.407$	710	1140
	$\underline{1}.329$	848	1069
	$\underline{1}.263$	988	989
	$\underline{1}.213$	1108	950
	$\underline{1}.184$	1186	916
0.667	$\underline{1}.460$	0.000627	1174
	$\underline{1}.395$	729	1117
	$\underline{1}.324$	857	1059
	$\underline{1}.247$	1026	983
	$\underline{1}.210$	1116	943
	$\underline{1}.178$	1200	912
0.725	$\underline{1}.473$	0.000609	1190
	$\underline{1}.454$	771	1100
	$\underline{1}.238$	1048	972
	1.176	1206	911

2. Theoretical

For reasons already explained above, it is unlikely that any complete theoretical treatment can be developed to cover all the cases of falling drops. An attempt has been made to deal with the case of the rigid sphere; this work was never satisfactorily completed but the preliminary attack shewed at least an approach to the true functional form.

The method used was as follows:- if a body in steady motion experiences a drag, then it must be possible to prove the existence of this drag by considering the flux of momentum through a surface surrounding the body. Such a body sets up a wake, the flow in which is laminar or turbulent according to the magnitude of the Reynolds' number. The wake thus acts as a diffusing agent for momentum, and since in addition the body plus its wake pushes the fluid away laterally, the flow set up in the main body of the fluid is similar to that caused by a source.

In the case of a sphere, the flow is similar to that generated by a point source, and using the equations developed at Porton for the diffusion of matter from a point source, it follows that in the case of a turbulent wake the drag coefficient is proportional to $R^{-1/4}$ where R is the Reynolds number. This relationship is reasonably verified by wind tunnel measurements for $10^2 < R < 3 \times 10^3$, but for $R > 3 \times 10^3$ the

For rigid spheres of moderate size it follows that the terminal velocity is given by an expression of the form

$$v = \left\{\frac{kg(\rho - \rho_1)}{\rho}\right\}^{4/7} \frac{d^{5/7}}{\nu^{1/7}}$$

where k is a dimensionless constant whose value is about 0.46

From this it follows that the variation of terminal velocity with height is governed by the term $\rho^{4/7} \nu^{-1/7}$ From experiments with liquids sprayed from aircraft it is thought that this indicates correctly the order of magnitude of the change of terminal velocity with height for small drops.

Certain empirical expressions, of the nature of interpolation formulae, have also been developed. These may be briefly summarized as follows:-

Experiments with various liquids have shown that drops falling through air behave, in all cases, like rigid spheres if $\frac{\rho_1 d^2 g}{T} < 0.4$

Under these circumstances the terminal velocities can be calculated from the following equation which is valid for spheres such that:-

$1.5 < \log_{10} R. < 4.2$ where $R = \frac{vd}{\nu}$ = Reynolds' number.

$$\log_{10} R = 2.655 \sqrt{\log_{10} R^2 - 0.0907} - 3.271 \qquad \text{.......(1)}$$

where ψ = drag coefficient = $\frac{(m - m')\,g}{\pi d^2 \rho_1 v^2}$.

m = mass of drop m' = mass of fluid displaced (usually negligible)

This equation holds for rigid spheres in fluids other than air.

If $\frac{\rho_1 d^2 g}{T}$ exceeds 0.4 the drops first experience a deformation in shape becoming flattended on the leading surface. Then, for larger drops, circulation inside the drop begins to affect the shape and the velocity. Finally, very large drops will accelerate to a critical speed and burst. This phenomenon is very beautiful to watch from above - the drop suddenly flies apart into a circlet of smaller drops possessing quite a remarkable symmetry.

The following empirical equation probably forecasts the terminal velocities of drops of all liquids falling in air down to half an atmosphere for $0.4 < \frac{\rho d^2 g}{T} < 1.4$, within about 3%. This is believed to represent the range of sizes over which circulation effects are negligible in all cases. For some liquids the equation is quite good up to much higher values of $\frac{\rho d^2 g}{T}$

$$\log_{10} R = 2.655 \sqrt{\frac{\log\sqrt{R^2} - f_1}{f_2}} - f_3 \quad \ldots\ldots\ldots(2)$$

where $f_1 = -0.460 + 1.012 \log_{10} \frac{\rho d^2 g}{T} + 0.225 \left(\log_{10} \frac{\rho d^2 g}{T}\right)^2$

$$f_2 = 0.933 - 0.167 \log_{10} \frac{\rho d^2 g}{T}$$

$$f_3 = 3.306 + 0.118 \log_{10} \frac{\rho d^2 g}{T} + 0.0765 \left(\log_{10} \frac{\rho d^2 g}{T}\right)^2$$

The equation is susceptible of improvement and is the subject of research at low priority at the moment.

True dynamical similarity between falling drop systems at the bottom of the 30 ft. tube is not always attained owing to the comparatively slow growth of circulation in large drops. The following experiments have an interesting bearing on this point.

An apparatus for making large drops was mounted at a certain height and drops of gradually increasing size allowed to fall to the ground. Up to a certain size all drops reached the ground intact, but, quite suddenly, at a critical size some of the drops began to burst in the last 10 ft. or so of their fall. Calling this critical diameter d_c the following experimental results are available.

Liquid	ρ_1	T	Viscosity of drop	$\frac{\rho_1 d_c^2 g}{T}$ (30ft)	$\frac{\rho_1 d_c^2 g}{T}$ (40ft)
Ethylene dibromide	2.17 g/cm³	38.1 dynes/cm.	0.016 poise	17.2	
Water	1 "	72.8 "	0.010 "	21.6	15.7
Methyl salicylate	1.18 "	38.6 "	0.033 "	26.5	18.3

Two points emerge from these figures:-

(a) The critical diameter for a given liquid is less for the greater height.

(b) The critical diameter at a given height increases with the kinematic viscosity of the liquid.

Hence it is of great importance that any subsequent experiments on the bursting of water drops be carried out from a sufficient height. It is also clear that the internal circulation of the drop is an important factor.

The equation given above predicts with roughly equal accuracy the velocities of 0.5 cm. diameter drops of methyl salicylate (viscosity 0.033 poise) and of methyl salicylate artificially thickened to 6 poise. On the other hand it overestimates the velocities of large water drops and drops of water made viscous with sugar. There is, therefore, some uncertainty as to whether the height of the apparatus was insufficient in the latter cases or whether circulation effects were more pronounced in the water drops which went up to larger diameters owing to the high surface tension. The former possibility is difficult to credit on account of the negligible acceleration which was detected and the latter is doubtful owing to the results for viscous (2 poise) water. Further improvement of the equation may elucidate this point since the exact incidence of circulation may be made definitely predictable.

On the experimental side apparatus has been developed for delivering drops which are both larger and smaller than those readily obtainable from simple tips.

II. Effect of Atmospheric Conditions on Falling Drops

(a) Evaporation

In meteorological problems connected with drops of water falling through the air it is justifiable to assume that

(i) the vapour pressure of the water vapour is small compared with atmospheric pressure,

(ii) water vapour behaves like a perfect gas,

(iii) free convection effects are negligible.

If, in addition we make the following assumptions, which are not quite as evident _a priori_,

(iv) saturation vapour pressure is maintained at the drop surface,

(v) the surface temperature of the drop is that of the ambient air,

then for drops whose shape is independent of velocity, the rate of evaporation is given by

$$E = \frac{2\pi D M (p_s - p_o) d}{RT} \cdot f\left(\frac{\upsilon d}{\nu}, \frac{D}{\nu}\right)$$

Here

E = rate of loss of weight, g/sec
D = molecular diffusivity, cm^2/sec
M = molecular weight of liquid,
p_s = saturation vapour pressure at temperature T, dynes/cm^2
p_o = vapour pressure of evaporating liquid in air at infinity, dynes/cm^2
d = drop diameter, cm.
R = gas constant, ergs/g.mol.
T = absolute temperature
υ = mean velocity of air past drop, cm/sec.
ν = kinematic viscosity of air, cm^2/sec.

In the case of still air, $f = 1$; this expression has been verified experimentally, and for various pure liquids the rate of change of surface area with time remains constant for a considerable range of drop diameter as demanded by the above equation.

In the case of a drop in relative motion to the air the technique employed was as follows:-

A drop, of about 1 mm. diameter, of the liquid under investigation was suspended on a fine platinum wire of diameter 0.05 mm. on the axis of a horizontal tube of cross-sectional area about 1 sq. cm. through which air could be passed at various reproducible rates of flow. In some experiments the drop was suspended from one junction of a thermocouple made of very fine wire. The temperature of the stream of air could be maintained at a fixed value by means of an arrangement whereby warm and cold air could be mixed together in varying proportions before passing into the tube in which the drop was suspended. A thermocouple, made of very fine wires, was arranged so that its junction was situated very near to the drop, on the leeward side of the latter. By manipulation of a tap controlling the mixing of the warm and cold air streams, the temperature of the air passing over the drop, as determined from the thermocouple reading, could be maintained at a constant value. Measurements of the horizontal diameter of the hanging drop were made from time to time during its exposure to the air stream, by means of a measuring microscope fitted with an adjustable filar micrometer-eyepiece. Since the shape of the drop was distorted by the airstream, the latter was always stopped while measuring the drop, which had then returned to its normal shape. Drops of the size used in these experiments were very nearly spherical when suspended in the manner described when the air-stream was stopped, but became rather pear-shaped when evaporation had considerably reduced their size. No measurements were taken after the drop had departed considerably from the truly spherical shape. The true velocities of the stream of air at the point in the tube occupied by the drop were measured, in a subsidiary series of experiments, by means of a very small pitot-tube inserted into the horizontal tube at the point normally occupied by the drop under investigation.

It has been found that the results of measurements made at Porton on a variety of liquids used in chemical warfare are in accordance with the theory developed by Frössling (Gerlands Beit., 52, 170, 1938). After making allowance for the cooling of the drop by evaporation Frössling found that for nitrobenzene, aniline, water and solid napthalene the velocity function f in the above expression could be well represented by

$$f = 1 + \frac{0.276}{}$$

where R = vd/ , the Reynolds number for a sphere of diameter d equal in volume to the drop.

The experiments covered Reynolds' numbers from 2 to 800 using drops up to about 1 mm. diameter held on fine suspensions. His equation has been found to apply to a number of chemical warfare agents indicating that if the drop temperature instead of the air temperature is used in obtaining the saturation vapour pressure the other assumptions can be relied upon to introduce only small errors for substances of moderate or low volatility.

Powell (Proc. Brit. Ass. Ref. 36.1.1939 - 40) has carried out experiments on evaporation from spheres of ice and water for Reynolds' numbers between 100 and 7000 usin spheres from 2 to 15.5 cm. diameter and measuring the actual surface temperatures. As would be anticipated from the absolute evaporation rates, which were only of the order of 1/300 of the maximum possible rate for water, no difference was found between ice and water. His results are inevitably given in terms of two dimensionless groups instead of the three of the complete equation above, since neither the diffusion constant nor the kinematic viscosity of air were varied. Nevertheless his figures can be correlated with Frösslings' equation, extrapolated to high values of the Reynolds' number, in the following manner.

Powell published his results in an equation which is not dimensionless

$$\frac{E\,D}{d\,(p_s - p_o)\text{mm}} = 2.83.10^{-8}\left(\frac{\upsilon d}{\nu}\right)^{0.62} \text{(pressures in mms.)}$$

The left hand team of this equation can be obtained from Frösslings' equation as follows

$$\frac{E\,D}{d(p_s - p_o)\text{mm.}} = \frac{2660\pi D^2 M}{R\;T}\left\{1 + \frac{0.276}{\sqrt[3]{D/\nu}}\quad\frac{\upsilon d}{\nu}\right\}$$

By substituting the following values for water in the latter equations a comparison between Powell and Frössling can be made.

$D = 0.22\ \text{cm}^2/\text{sec}$ at 760 mm. and 273°K

Table of values of $\frac{E\,D}{d(p_s - p_o)\text{mm.}}$

R =	100	500	1000	5000	10000
Powell	0.53	1.43	2.20	5.94	$9.16.10^{-6}$ cm²
Frössling	1.04	1.94	2.61	5.45	$7.56.10^{-6}$ cm²

Bearing in mind the very different scale of the experiments and also that Frössling is probably most and Powell least accurate at low Reynolds' numbers, these figures must be regarded as being in remarkable agreement and suggest that Frössling's equation may be used up to higher values of the Reynolds' number than was originally intended.

The situation regarding evaporation may therefore be summed up by saying that provided there is no serious di distortion from the spherical shape, the Frössling equation gives a very close approximation to the actual rate of loss. Even for large drops, it is probable that the Frössling formula is correct as regards order of magnitude.

(b) The temperature of Falling Drops

The main work on this problem has been theoretical and has been confined to liquids used in chemical warfare.

In the mathematical analysis it was first assumed that

(i) the drops are always spherical

(ii) the flow of heat inside the drop is according to the equation of molecular conduction of heat

(iii) the isothermal surfaces inside the drop are concentric spheres.

(iv) radiation and evaporation have no effect

(v) the surface of the drop is at the temperature of the surrounding air.

If r = distance from the centre of the drop,
a = radius of the drop,
t = time,
$\theta(r)$ = temperature throughout the drop,
$\bar{\theta}$ = mean temperature of the drop,
k = conductivity of the liquid forming the drop (cm^2/sec)

then the problem requires the solution of

$$\frac{\partial\theta}{\partial t} = k\left(\frac{\partial^2\theta}{\partial r^2} + \frac{2}{r}\frac{\partial\theta}{\partial r}\right.$$

subject to the boundary conditions,

$$\theta = 0 \text{ for } t = 0 \text{ , } 0 \leq r \leq a$$

$$\theta = f(t) \text{ " } r = a \text{ , } t > 0$$

This implies that the initial air temperature is taken as zero, and if it is further assumed that there is a constant lapse rate equal to α degrees per second of fall, then the temperature of the surface of the drop is αt at a time t seconds after zero.

It may then be shewn that

$$\bar{\theta} = \frac{3}{a^3}\int_0^a \theta r^2 dr = \alpha t - \frac{\alpha a^2}{15k} + \frac{6\alpha a^2}{k\pi^4}\sum_{n=1}^{\infty} n^{-4} \exp\left(-\frac{kn^2\pi^2 t}{a^2}\right)$$

Hence a drop formed at air temperature would be $\alpha a^2/15k$ degrees colder than the surrounding air after a prolonged fall, in the absence of radiation and evaporation effects. This lag is generally very small; for a 2 mm. diameter drop of liquid for which $k = 10^{-2} cm^2/sec$ falling through air whose lapse rate is 2/3 of the dry adiabatic the lag is only about 0.003°C.

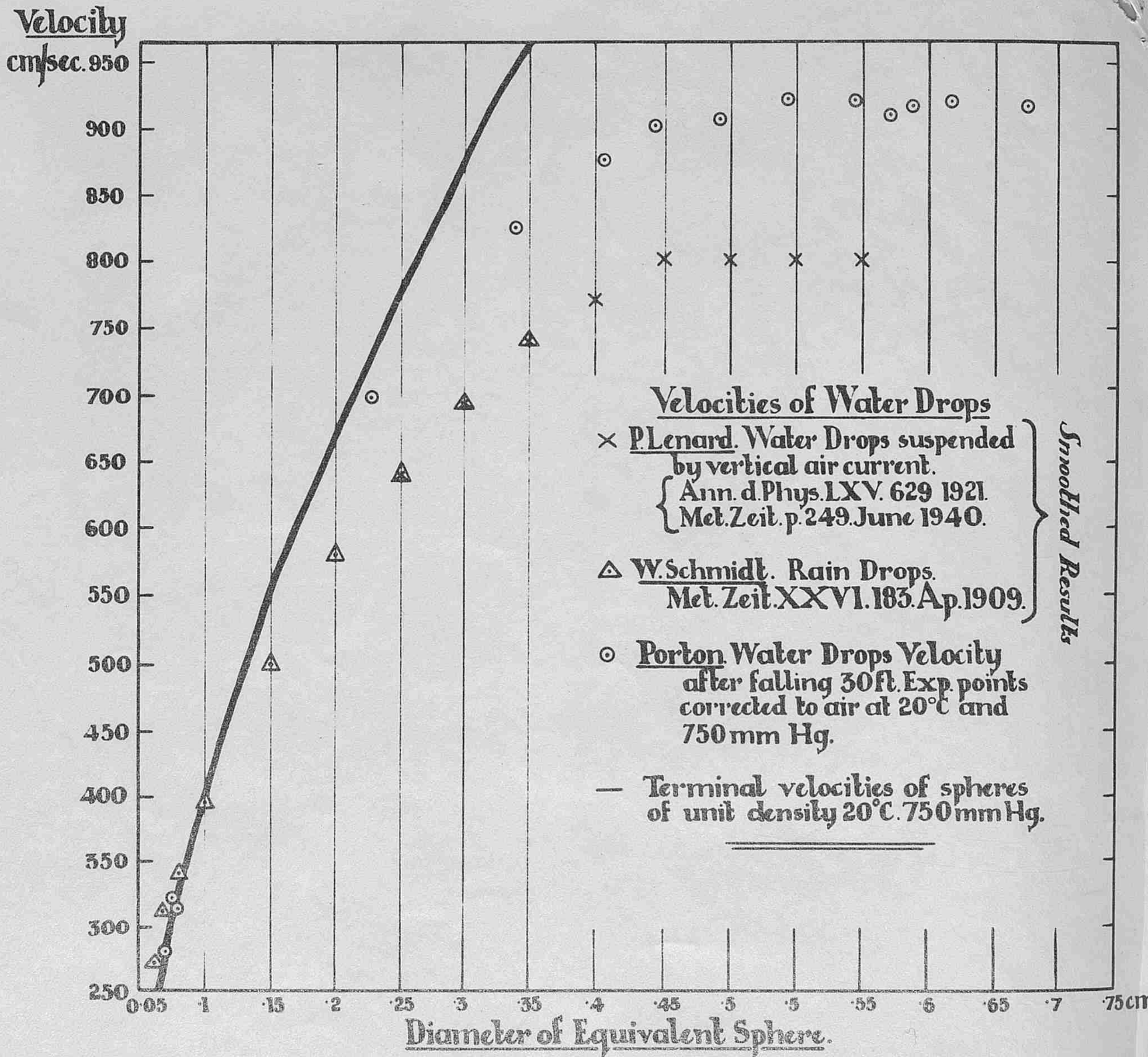
It is difficult to include the self cooling due to evaporation in the preceding calulations, but an estimate of the order of magnitude of this effect was obtained as follows:- if H is the amount of heat lost by the drop in evaporating in time t, then

$$\frac{dH}{dt} = \frac{\rho_1 L}{4\sqrt{\pi}}\sqrt{S}\,\frac{dS}{dt}$$

where S is the surface of the drop and L is the latent heat of vapourization. In the case of the liquids used in chemical warfare (of low vapour pressure and whose latent heat of vapourization is about one tenth of that of water) it was calculated that the rate of loss of heat due to the slow evaporation was small compared with the rate of gain of heat by conduction from the air through which the drop falls, but the case of water drops is very different. It is in fact obvious that the mean temperature of a rain drop should be about the same as that of the wet bulb temperature of the surrounding air.

A later investigation attempted to allow for the effects of radiation, and again it was concluded that for chemical warfare liquids the drops behave effectively like well ventilated thermometers of small lag and as such are never very far from air temperature. These conclusions were more or less verified by observations on sprays of nitrobenzene from very great heights.

The Porton investigations throw little or no light on the complex problem of the formation and precipitation of hail in air whose surface temperature is above freezing point. One of the main difficulties is the lack of knowledge of the conducitivity of the hailstone.


Velocity
cm/sec.
950
900
850
800
750
700
650
600
550
500
450
400
350
300
250
0·05
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·55
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·7
·75cm
Diameter of Equivalent Sphere.
Velocities of Water Drops
× P. Lenard. Water Drops suspended by vertical air current.
Ann. d. Phys. LXV. 629 1921.
Met. Zeit. p. 249. June 1940.
△ W. Schmidt. Rain Drops. Met. Zeit. XXVI. 183. Ap. 1909.
Smoothed Results
⊙ Porton. Water Drops Velocity after falling 30 ft. Exp. points corrected to air at 20°C and 750 mm Hg.
— Terminal velocities of spheres of unit density 20°C. 750 mm Hg.

AIR MINISTRY

M.R.P.43
12th August 1942

METEOROLOGICAL RESEARCH COMMITTEE

Note on the Use of the Richardson Number in Meteorlogical Problems

by O. G. Sutton

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Introduction

The Richardson Number

$$Ri = \frac{g(\frac{\partial T}{\partial z} + \Gamma)}{T(\frac{\partial u}{\partial z})^2} \qquad \ldots\ldots\ldots \quad (1)$$

JUL 3 1944

where T = absolute temperature at height z

Γ = adiabatic lapse rate,

u = mean wind velocity at height z,

has been suggested as the natural counterpart of the Reynolds number R in problems of turbulence in a medium stratified by heat. The object of the present Note is to give certain results which have been obtained in investigations on the diffusion of smoke clouds and which tend to support this hypothesis.

The Richardson number as the criterion of turbulence

Prandtl's (1) derivation of the Richardson number is briefly as follows:- suppose we have a fluid moving in a horizontal direction with a stratification of density (ρ) in the vertical direction. It is supposed that the fluid is in a state of turbulence and that there is a gradient of mean velocity (u) in the vertical. The work done against gravity in lifting the heavier masses and depressing the lighter ones over a length l is proportional to $gl_2 d\rho/dz$. This work must be done at the expense of the energy of supply of the turbulent mixing which is proportional to $\tau l\ du/dz$ where l is the Prandtl mixing length and τ is the eddy shearing stress. According to Prandtl

$$\tau = \rho\overline{u_1 v_1} = k\rho l \frac{du}{dz}$$

where k is an undertermined correlation factor so that the work done by the 'eddy friction' is $k\varrho l_2(du/dz)_2$ Hence, whether or not turbulence tends to grow or to be damped out must depend on the relative magnitudes of $gl_2 d\varrho/dz$ and $k\varrho l^2(du/dz)_2$ i.e. the criterion of turbulence in a stratified medium must depend upon the dimensionless quantity

$$Ri = \frac{g}{\varrho}\frac{d\varrho}{dz} \Big/ \left(\frac{du}{dz}\right)_2 \qquad \ldots\ldots\ldots\ldots\ (2)$$

In a compressible fluid this quantity takes the form (1) above.

Prandtl suggested that the limit of stability is given by Ri = 2 but G. I. Taylor (2) later shewed this to be incorrect and that the argument should give Ri = 1 in agreement with Richardson's original derivation (3). Later Taylor (4) and Goldstein (5) examined the stability of plane laminar flow of an ideal non compressible fluid with a stable density distribution, using the method of small oscillations. For mathematical simplicity the investigations were limited to cases in which the velocity profiles consisted of segments of straight lines. The results of these investigations were expressed in terms of two non dimensional quantities Ri and λ/δ, where λ is the wave length of the disturbances and δ is a characteristic length associated with the velocity profile (boundary layer thickness). Most of the cases investigated give no proper stability limit for Ri, but for every value of Ri there is always a definite, if narrow, region of unstable disturbance wave length λ.

Taylor found a definite result in the case of a fluid extending to infinity either upwards or downwards with a linear velocity profile and a continuous density distribution. He gave Ri = 1/4 as the stability limit. The same value was found by Goldstein for a very similar case.

The so called "Hot-Cold" Wind Tunnel at Göttingen provided an opportunity of making observations on the quenching of turbulence by a stable stratification. This tunnel is 16 m. long, 1 m. wide and 25 cm. high. Reichardt, who made velocity and temperature explorations in this tunnel, confined himself to the stable case, in which the upper surface was heated by steam (100°C) and the lower cooled by running water (10°C), thus giving the enormous 'inversion' of 90°C over 25 cm. The Taylor-Goldstein criterion is not applicable to this case, since the velocity distribution was not linear.

A very extensive mathematical investigation was made by Schlichting (6) who used a curved profile and included the effects of friction. In his analysis, Schlichting found it necessary to introduce the so called 'Froude number' $F = u_s/\delta g$, and he determined a critical number R(Ri,F) and thereby a critical Richardson number (Ri) crit. i.e. the largest value of Ri which an undamped disturbance wave length exists in the profiles. Tollmien (7) had previously shewn that for the flow of a homogeneous fluid over a plate the critical Reynolds number is 575. Schlichting found that the value of (Ri) crit. depended somewhat on the Froude number; for F = 0 (the case in which the inertia effect of the stratification is negligible compared with purely gravitational effects) the value of (Ri) crit. is 0.0409 i.e. about 1/24, while for F = 5, the value fell to 0.029. These values are considerably below the value Ri = 0.25 found by Taylor and Goldstein in their work on perfect fluids and linear profiles. Reichardt's measurements shew excellent agreement with Schlichting's theoretical deductions, and indicate that for $R > 575$, turbulence appears if $Ri < 1/24$ and is damped out if $Ri > 1/24$.

In meteorological literature, two contributions have appeared in recent years by Paeschke (8) and Durst (9) respectively. Paeschke examined velocity and temperature profiles near the surface and his results appear to shew a measure of agreement with the Schlichting criterion (Ri) crit. = 1/24. Durst, on the other hand, by considering the phenomenon of the building up and break down of the noctunial surface inversion gives good reason for believing that the original Richardson criterion (Ri) crit. = 1 may hold in the lower atmosphere.

(It may be mentioned here that it is doubtful if true laminar flow ever appears, except perhaps momentarily, in the lower atmosphere. Best (10) found only one case in which a linear velocity profile existed near the ground, so that a sharp criterion is hardly to be expected).

The Richardson number as a universal parameter

The above investigations deal with the Richardson number as the criterion for the generation of turbulence. There is, however, another reason for interest in this quantity. In aerodynamical problems, the Reynolds number is the natural universal parameter (e.g. the drag coefficients of spheres plotted against R all fall on one curve, no matter what the size of the sphere or the velocity of the fluid).

problems involving stratified flow the Richardson number should play the same part of a universal parameter.

There is evidence for this in measurements of the rate of diffusion of smoke clouds in the lower atmosphere. In the work described below only the lateral (crosswind) component of diffusion is considered. The fact that this quantity appears to be uniquely determined by the Richardson number which involves only vertical gradients may be accepted as an indication of the fact, first noticed by G. I. Taylor, that the lateral and vertical components of eddy diffusion tend to decrease simultaneously in the presence of a stabilizing gradient. The importance of the determination of a true universal parameter, analogous to the Reynolds number, for meteorological investigations is obvious; one has only to think of the difficulties of dealing with aerodynamical results if no such parameter as the Reynolds number existed to appreciate the chaotic state which would result.

Smoke Cloud Measurements (Porton Report No.1423)

In these trials, which were carried out over flat country at the Royal Airship Establishment, Cardington, smoke generators were functioned at a point and the resulting clouds sampled 100 m. downwind. During the trials the velocity profile was determined by small vane type airmeters mounted on masts at heights of 1 m., 2 m. and 5 m. above the ground. The vertical gradient of temperature was measured as the difference in temperature given by two platinum resistance thermometers at heights of 1.2 m. and 7 m. The resistance elements were in aspirated anti-radiation housings and records were taken on a thread recording galvonometer. The actual temperature at 1.2 m. was measured by an aspirated thermometer.

Before discussing the results it is necessary to explain some of the terms used in chemical warfare. The cloud from a point source spreads out in the form of a cone, and at any fixed distance downwind the crosswind concentration curve is of the "cocked hat" variety and approximates to the normal curve of errors. The shape and dimensions of this curve depend upon the period of sampling, but for periods upwards of 2 minutes or so, the variation is slow. In these trials, the period of sampling was invariably 3 minutes.

The "width" of the cloud at a fixed distance downwind is defined as the distance between points on each side of the axis at which the concentration falls to 1/10 of the peak concentration.

problems involving stratified flow the Richardson number

The use of the width, as defined above, as the measurement of lateral diffusion has certain advantages. In the first place, since only the ratio of concentrations is involved, it is clear that the width sodefined is independent of the strength of the source. Secondly, it may be shewn theoretically and has been verified by trial that within limits, for any steady condition of density gradient, the width of a cloud varies only slowly with changes in wind velocity. The width of the cloud at 100 m. is therefore a very suitable indicator of the magnitude of lateral (crosswind) diffusion.

The relation between the widths of the clouds and the value of the Richardson number Ri is shown in the attached graph. The vertical gradient of velocity was measured as a finite difference ratio by using the velocities at 2 m. and 5 m., and the value of the temperature gradient was computed also as a finite difference ratio from the measured differences 1.2 m - 7 m.

It is obvious that there is a high correlation between the width and Ri and the best curve to fit the points is in fact given by

$$\text{width} = 46 - 60\sqrt{Ri},$$

this empirical equation applying only to positive values of Ri.

The above curve is included in the figure and it is seen that it fits the points well for values of Ri below 0.25 but for higher Ri, the width is underestimated. This poor fit for the high values of Ri makes it impossible to determine the critical value of Ri (if such exist) for which lateral diffusion is negligible, i.e. for which turbulence dies out. It may be noticed that for zero width, the above empirical equation yields Ri = 0.59, which suggests a measure of agreement with Durst's result. The same widths when plotted against the temperature gradient alone shewed a very wide scatter.

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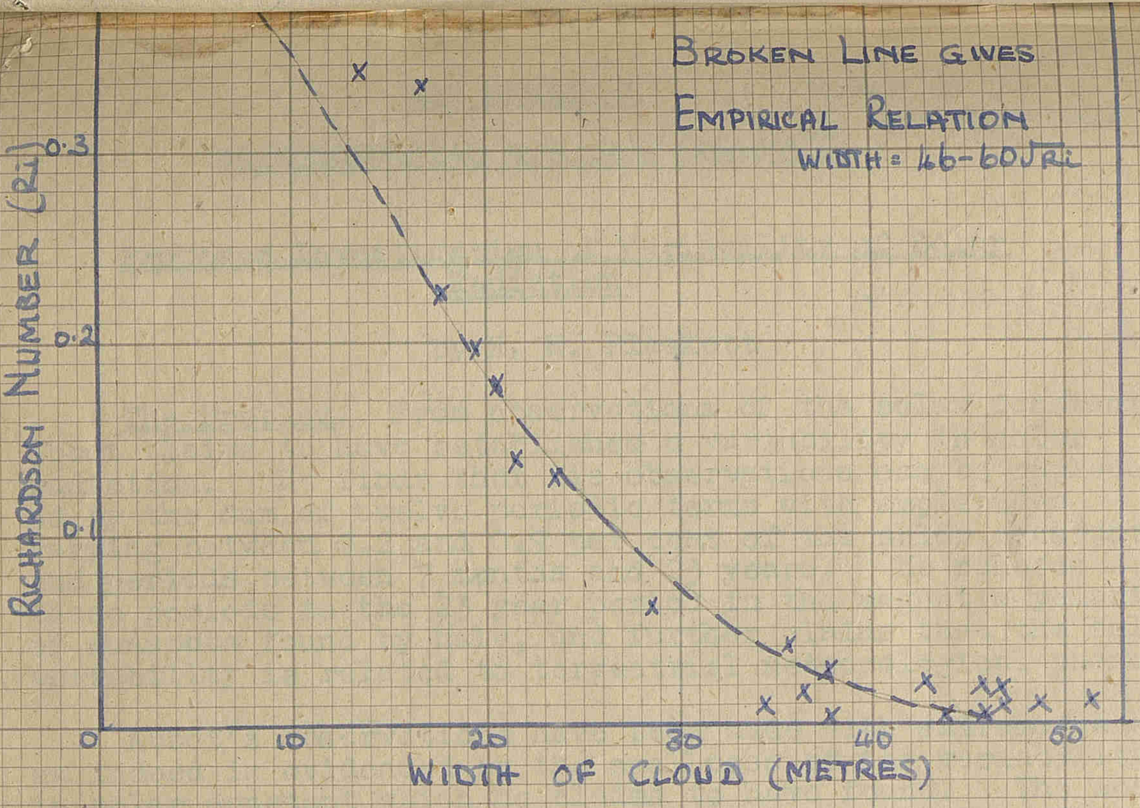

VARIATION OF WIDTH OF CLOUD WITH RICHARDSON NUMBER.
(21 EMISSIONS)

observing station (Montezuma, height 8900 feet), this correction is approximately one fourth of the derived solar constant, it is evident that any uncertainty in the correction will re-appear with considerable weight in that constant. The uncertainty in the transmission correction found by Abbot arises in two ways. In the first place, since his primary measurement is one of total radiation by the pyrheliometer, the theoretically exact method of finding the correction is a somewhat complex one in which the propagated error may be far from negligible; that this must be the case readily follows when it is recalled that the so called long method involves spectrobolometric measures of energy distribution in the solar spectrum at several zenith distances, determination of the transmission of the spectrobolometer and auxiliary apparatus as a function of wave length, estimation of water vapour absorption and estimation of the contributions to the solar constant of the ends of the spectrum beyond the range of the spectrobolometer. In the second place, and more serious, the transmission of the earth's atmosphere may not remain constant over the interval, somewhat greater than two hours, covered by Abbot's observations; a slow

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P 71

AIR MINISTRY

M.R.P.63
31 December 1942

METEOROLOGICAL RESEARCH COMMITTEE

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Proposed Investigation of Variability of Solar Radiation.

by H. H. Plaskett

The Problem. From observations of the solar constant Abbot claims to have established a variability in solar radiation. He finds that his observations can be represented by the sum of twelve periodic terms with amplitudes varying between 0.2 and 0.9 per cent of the solar constant, and periods between 7 months and 23 years. He relates these periods to the 23 year period in the magnetic polarity of sunspots found by Hale, and further claims that the solar variability is related to terrestrial meteorological phenomena.

Abbot's observations have been made with extreme care, as indeed is evidenced by the smallness of the variability in solar constant which he claims to have found. The most serious criticism to which the observations are open lies in the determination of the correction for the transmission of the earth's atmosphere. Since at his highest and best observing station (Montezuma, height 8900 feet), this correction is approximately one fourth of the derived solar constant, it is evident that any uncertainty in the correction will re-appear with considerable weight in that constant. The uncertainty in the transmission correction found by Abbot arises in two ways. In the first place, since his primary measurement is one of total radiation by the pyrheliometer, the theoretically exact method of finding the correction is a somewhat complex one in which the propagated error may be far from negligible; that this must be the case readily follows when it is recalled that the so called long method involves spectrobolometric measures of energy distribution in the solar spectrum at several zenith distances, determination of the transmission of the spectrobolometer and auxiliary apparatus as a function of wave length, estimation of water vapour absorption and estimation of the contributions to the solar constant of the ends of the spectrum beyond the range of the spectrobolometer. In the second place, and more serious, the transmission of the earth's atmosphere may not remain constant over the interval, somewhat greater than two hours, covered by Abbot's observations; a slow secular change of transmission cannot readily be detected from the observations and appears with its full weight as an error in the determination of the transmission.

Abbot strives to overcome these two difficulties, the seriousness of which he fully recognizes, by a so called short method of reduction. This method is based on the recognition by Abbot of a correlation between the brightness of the sky near the sun and the simultaneously measured atmospheric transmission coefficients; as applied in practice the sky brightness is only used as a correction to the precipitable water in the atmosphere to form a "function" which is correlated with the atmospheric transmission. Since the "function" is wholly empirical with little or no physical justification, and since indeed the method is based on the not wholly satisfactory long method of finding the correction for atmospheric transmission, it is clear that the short method affords no more satisfactory a means of finding the correction than does the theoretically justifiable long method.

In a previous memorandum to the Meteorological Research Committee (Paper M.R.P.31, dated 21.5.42), it was suggested that these difficulties in the determination of atmospheric transmission could be minimized by a departure from Abbot's procedure. Since the proposed new method would automatically yield the surface brightness of the sun outside the earth's atmosphere at several wave lengths, it is clear that measures by this method could be used to test the variability of solar radiation predicted by Abbot.

The proposed Method. The monochromatic surface brightness of the sun is to be determined at several wave lengths by a comparison with two radiation standards. One of these is a vacuum ribbon-filament lamp standardized by the National Physical Laboratory and the Nela Laboratories for colour temperature and brightness temperature, while the other is a special ribbon-filament lamp designed and calibrated at the Utrecht Physical Laboratory to give the surface brightness in ergs cm.$^{-2}$sec.$^{-1}$ as a function of wave length and current. The comparison is to be made from a Boston aircraft flown at a constant height of 25,000 feet, the comparison being made by visual spectrophotometric methods using a small ribbon filament lamp as a secondary standard. The determination of monochromatic surface brightness of the sun at several zenith distances by comparison with a radiation standard eliminates the necessity for finding the transmission of the optical system, and at the same time provides a wholly self-contained method of determining the transmission coefficient and the surface brightness of the sun outside the earth's atmosphere. Thus the proposed method eliminates at one stroke the long and involved process with its consequent error required by Abbot to find the solar constant outside the earth's atmosphere.

The more seriouserror in the correction for atmospheric transmission, which results from a change of transmission with time and occurs presumably as a result of variable amounts of dust carried into the air, is considerably reduced by making observations at a height of 25,000 feet. Not only at this height does turbulent mixing, limited to heights within a few thousand feet of the immediate ground level, play no part, but also convective mixing, resulting from heating of the ground by solar radiation, is much reduced above 25,000 feet where the lapse rate is small. If flights could only be carried out above the tropopause, unfortunately normally well above the practical ceiling of the Boston, any changes with time in atmospheric transmission due to air transport of dust, would of course be completely eliminated. However even at a height of 25,000 feet these changes should be small, and such change as remains could be minimized by making observations in the afternoon, when ground temperature is stabilized and shows a much smaller variation with time than in the morning hours.

A great advantage in making observations at a height of 25,000 feet arises from the fact that the air mass above the observer is less than six tenths of that over an observer at a station such as Montezuma with a height of 8900 feet. Consequently any error in the determination of the correction for atmospheric transmission appears with a correspondingly reduced wieght in the determination of solar surface-brightness.

Progress to Date. In the previous memorandum (M.R.P.31) it was suggested that the proposed investigation should be carried out jointly by the Meteorological Research Committee and the University Observatory, Oxford, the part of the latter being to supply the necessary equipment. At a meeting of the Committee on 25 June 1942 (letter of Dr. G. M. B. Dobson, dated 26.6.42) it was agreed that work in preparation for actual observations could be commenced informally, leaving for decision at a later date the assignment of the problem to a formal research programme.

In accordance with this informal agreement a Hilger Constant Deviation Spectroscope has been purchased by the University Observatory. A photometer unit for use with this instrument has been designed, closely following proposals by Buckley and Brookes (Jour. Sc. Inst., vol. 7, 305, 1930), and has been constructed by Casella and Co. In addition a heavy oak frame to carry the spectroscope and the photometer as a rigid unit has been constructed at the Observatory, and has been mounted with Silentbloc bearings (to minimize vibration) on a stand. This stand permits observation of the sun at various altitudes, and can itself be clamped to a table, to be built on the semi-circular rail in the rear gunner's cockpit of the Boston.

Since it is proposed to make observations of the sun between zenith distances of 70° and 85° (corresponding air masses 2.92 and 11.47), and since this interval is covered in less than one hour forty minutes for positive declinations of the sun, it is clear that observations should be made between March 21 and September 21. Further for the year 1943 Abbot (Annals Astroph. Obs. Smithsonian Inst., vol. 6, 190, 1942) predicts the following values of the solar constant:-

March	1.943	June	1.948
April	1.942	July	1.946
May	1.943	August	1.945
		September	1.945

These predicted differences are probably just large enough to permit of a test by the proposed method, particularly since for observations in the blue region, Abbot (loc. cit. p.166) predicts a percentage variability in surface brightness twice that shown by the solar constant.

Proposals for Future Work. In view of the stage now reached in the preparation of the equipment for the investigation, and in view of the fact that definitive observations could and should commence in the latter half of March, it is desirable that this problem should now be made part of a definite research programme. Since the observations would be made by the writer, it is desirable that the problem be made part of the research programme of the Navigation Section, A. and A.E.E., Boscombe Down, and be described as an investigation to be carried out for the

Meteorological Research Committee. A recommendation to this effect from the Committee would presumably lead to the necessary action by D.S.R.

The flying needed for this problem would be limited to two-hour flights in the late afternoon (between 1730 and 1930 B.S.T. at the end of April and in mid-August, and between 1920 and 2120 D.B.S.T. in mid-June). The number of flights would be limited by weather conditions, and would probably vary between 4 and 10 per month. It is highly probable that this limited number of flights could be made as required in one of the Bostons of the Met. Research Flight. The apparatus could be mounted so that it could in no way interfere with the meteorological investigations being carried out on the same aeroplane.

Before any flights can be made it will be necessary carefully to line up and test on the ground the spectrophotometric equipment now assembled at the University Observatory. In particular, considerable experimenting will have to be done with the mounting of the standard lamps for the ground standardization. It is suggested that this work be done at Oxford where the necessary facilities for testing and for making modifications are more readily available than at Boscombe Down. Provided the Meteorological Research Committee agrees to these various proposals this testing could be carried out during a week early in February. On its completion the apparatus would be brought to Boscombe Down for preliminary flight trials, and for the proposed set of observations.

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AIR MINISTRY

M.R.P. 64
12 January 1943

METEOROLOGICAL RESEARCH COMMITTEE

Measurement of Turbulence in the Free Atmosphere.

by
C.S. Durst.

During the course of conversation in August 1941, between Dr.Hopkins of the N.P.L., Mr.A.C.Best and the writer, Dr.Hopkins stated that he believed the swing of the can of a radio wind-finding balloon might give a measure of the disturbance due to eddies and he also stated that the swing of the can was measured in the early ascents observed with a loop aerial.

To test how far the observations showed real swinging of the can, some of the early records were examined and it was found that the magnitudes of the simultaneous swings observed at two independent stations were closely correlated on some occasions, though on others there appeared to be considerable differences which were believed to be due to some observers being less expert than others.

On one occasion when two ascents were made within 1¼ hours of each other the swing appeared to die down at about the same level on both occasions and to increase to a maximum at about 22,000 feet in both ascents.

These results seemed to justify further observations being made, and a loop aerial was borrowed from the N.P.L. and erected at Larkhill. Various delays occured and the first observation of the new series was made on 7th July, 1942. Since then 8 other observations have been obtained giving 9 in all, though of these the first two gave less reliable results than the later ones as observation technique was improved after 8th July.

It has been the intention to get the observations made in typical and varied synoptic situations and to some extent this has been achieved.

It was suggested that if the swing of the can was due to the amount of eddying in the atmosphere and the amount of eddying was related to Richardson's criterion:

$$\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2 - \frac{g}{T} \cdot \left(\frac{\partial T}{\partial z} + \Gamma\right) \quad +ve \text{ or } -ve \qquad \ldots.\ 1.$$

then a plot of the magnitude of the swing alongside this expression would serve as a verification of the reality of the swing as a representation of the amount of edying and of the truth that Richardson's criterion gave, in fact, a measure of the stability of the atmosphere. Accordingly in each of the nine cases a diagram has been drawn showing the curve of swing and the curve of expression (1). These diagrams are given as figures 1 to 9. Brief notes as to the synoptic situation and the characters of the curves are given below:

July 7, 1942. Maritime polar air circulating around a stale low pressure centre north-west of Scotland. No fronts in neighbourhood of British Isles.

There is a pronounced increase in the swinging observed at and above the tropopause, where stability as shown by the Richardson criterion is great. Above this the swinging decreased abruptly and decidedly.

July 8, 1942. Maritime polar air circulating around a stale low pressure centre north-west of Scotland. A cold front moving south-east was about 300 miles north-west of Larkhill.

There is again a pronounced increase in the swinging observed at and above the tropopause. Above 160 mbs. there is a sudden decrease in swing.

July 23, 1942. Maritime polar air near surface and probably extending to 10,000 to 12,000 feet (600 to 700 mbs.); warm air probably lying above that height. A cold front cut the surface about 200 miles south-east of Larkhill.

ENCLOSURE 1 TO IR-283-51
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The swing increases fairly steadily between 600 mbs. and 350 mbs. and then decreases. There is a sudden change in the stability curve between 575 and 525 mbs. and another about 350 mbs.

July 24, 1942. In an active warm sector to a depression moving over N.W. Scotland. The swing increased near the tropopause.

August 10, 1942. Just in front of an active warm front to an occluded depression moving north-east to the west of Scotland. Ascent was in the warm air.

The swing was not measured below 650 mbs. The amount of swing was rather high. There was a decrease in swing below the tropopause and an increase at and above it.

August 13, 1942. Just in front of an active warm front to an occluded depression moving north-east to the south-west of Iceland. Ascent was in the warm air except in perhaps the first 2,000 feet.

The swing increased from 800 mbs. to 550 mbs. where stability was fairly great. The swing increased just above the tropopause but decreased suddenly at about 250 mbs. The curve of stability and of swing are almost exactly opposed from 800 mbs. up to the tropopause.

August 28, 1942. A depression centred over Bay of Biscay. Easterly winds at the surface over British Isles, fronts not affecting air over Larkhill.

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The swing is comparatively small at all heights, perhaps it increases above the tropopause.

September 1, 1942. A cold occlusion associated with a depression near Iceland was about 100 miles west of Larkhill.

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It is remarkable that on several occasions the curves of variation of swing with height are almost exactly the mirror of the curves of the expression

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The outstanding feature of the curves of amplitude of swing is the decided tendency for swing to increase in the neighbourhood and for some distance above the tropopause. On a number of occasions there is a sharp decrease in amplitude some distance above the tropopause. In the one case in which Richardson's criterion was fulfilled, July 23rd, the swing was greater than during any of the other ascents (the swing plotted for July 7th and 8th was measured in a somewhat different way and is not comparable) and the maximum swing is just above the region where stability appears to have broken down.

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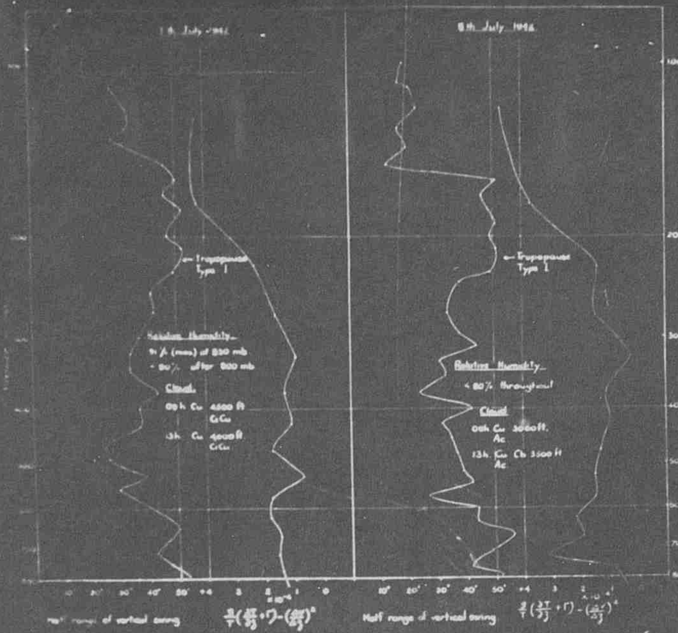
Larkhill



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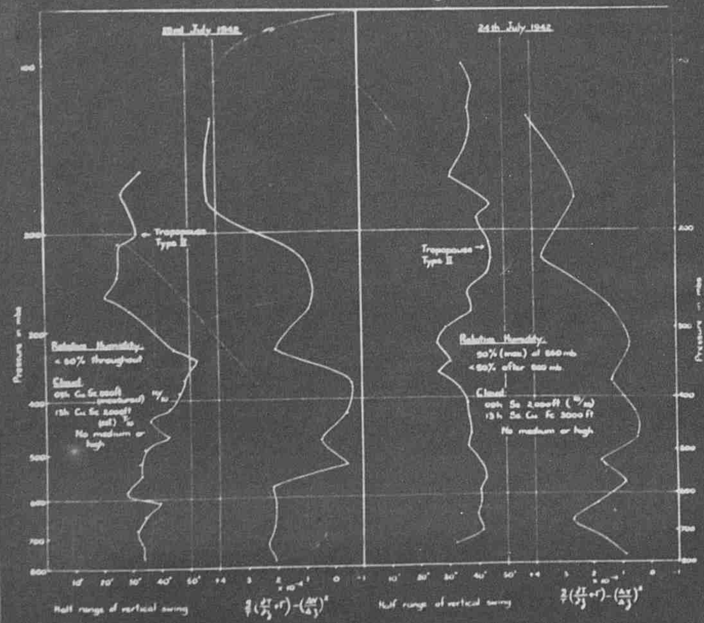
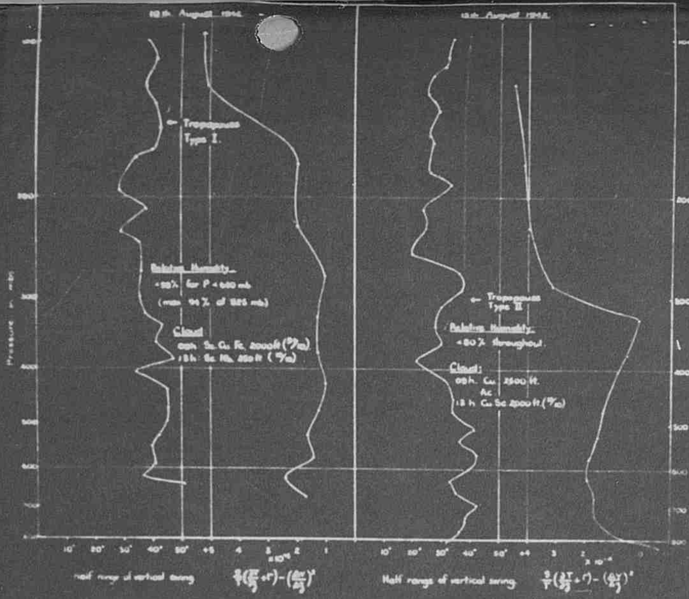
Larkhill





Larkhill

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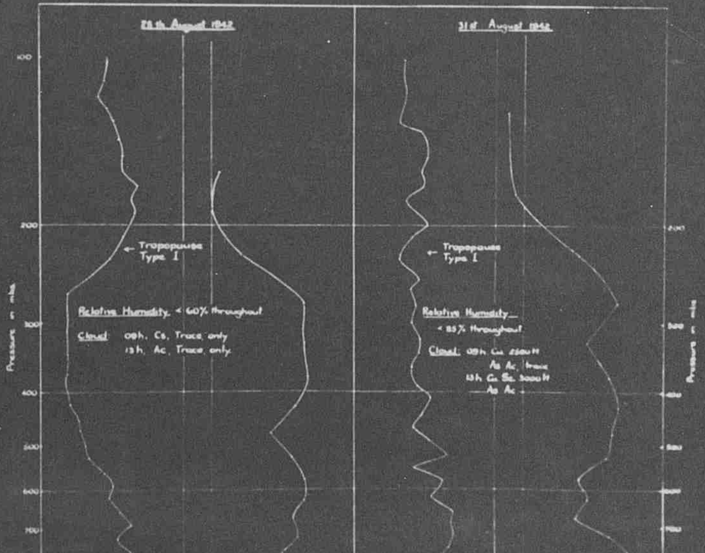
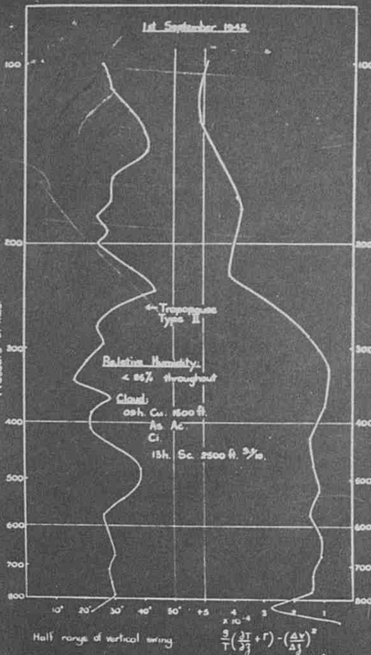


Larkhill



1st September 1942

← Tropopause Type II

Relative Humidity:
< 86% throughout

Cloud:
09h. Cu. 1500 ft.
As. Ac.
Ci.
13h. Sc. 2500 ft. 3/10.

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AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

M.R.P.64
12 January 1943

Measurement of Turbulence in the Free Atmosphere.

by
C.S. Durst.

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by Ape:

AIR MINISTRY

M.R.P.68
1 February 1943

METEOROLOGICAL RESEARCH COMMITTEE

THE INFLUENCE OF RADIATION ON CLOUD SHEETS.+

By Professor D. Brunt, Sc.D., F.R.S.

It has long been realised that a sheet of cloud will gain heat at its base by absorption of upward radiation from the ground, and will lose heat by radiation from its upper surface. This must occur, no matter what the detailed structure of the water vapour spectrum may be, since the temperature of the air decreases with height above the ground, and since there are bands of wavelengths to which water vapour is transparent, or nearly transparent.

Particularly at night, when no absorption of solar radiation is possible, cloud will therefore tend to become unstable, by the increase of lapse-rate within it. This instability will be most marked where the sum of the rate of gain of heat at the base and of the rate of loss at the upper surface is greatest. A rough computation indicates that this is about the Cirro-cumulus level (see Durst, 1933). Even at other levels there will be a marked tendency to produce instability, leading to the formation of typical instability patterns in the cloud (Brunt, 1937). If, for example, the wind speed increases with height, within the range of height where the cloud occurs, the cloud will break up into long rolls, which will be transverse to the wind direction if the rate of increase of wind with height is slight, and parallel to the wind direction if the rate of increase of wind with height is great. This phenomenon of the breaking-up of a cloud sheet into rolls, as a first step towards clearing, is frequently observed after sunset. The actual dissipation of the cloud may be attributed to two causes; first, the mixing of the cloudy air with the surrounding clear air, as a result of the motion in the rolls, and secondly, the evaporation due to general subsidence of the cloud, and the consequent rise of temperature.

+ This paper formed one section of a paper entitled 'Radiation in the Atmosphere', read at the Toronto meeting of the Royal Meteorological Society, Aug.28, 1939, but not included in the paper as published in Toronto Supplement to the Quarterly Journal, 1940.

When the cloud is at medium or low levels, the loss of heat from the top is greater than the gain of heat at the base. Thus the cloud suffers a net loss of heat, its average temperature is decreased, and its average density is increased. It will consequently tend to subside through its environment. A small cloud sheet can subside bodily through its environment, but an extensive cloud sheet cannot subside bodily without change of structure. It can only descend piecemeal, its descent being akin to turbulent motion rather than to steady motion in the vertical.

In the diagram, curve (1), let AB represent on a tephigram the distribution of temperature in the vertical, the cloud being at the level of B, the lapse-rate below the cloud being intermediate between the dry and saturated adiabatic rates. The cloud is then unstable, and any disturbance will cause it to descend, its representtative point moving down the saturated adibatic through B, either until it reaches C on the curve of the environment, or, if the cloud is completely dissipated before reaching C, say at C^1, the resulting clear air formed by the dissipation of the cloud will descend to D.

When the lapse-rate is less than the saturated adiabatic, as in curve (2), the cloud will first cool without change of level, and the representative point will first move along the isobar through B to a point C to the left of B, and will then descend along the saturated adiabatic through C to the point D. This process will be repeated until the cloud has dissipated. It will be seen that at D more water vapour is required to saturate a given mass of dry air than at B. Hence some of the liquid water present in the cloud is evaporated during the passage from B to D. The process of descent will be continued until the cloud has dissipated.

When the lapse-rate changes its sign, and there is an inversion below the cloud as in curve (3), the descent of the cloud involves an increase in the density of the cloud, since less water vapour is required to saturate the air at D than at B, which results in further condensation of water vapour into liquid water during the passage from B to D.

It is seen that the limiting lapse-rate through which a cloud can descend without further condensation or evaporation of liquid is that corresponding to the slope of the dew-point lines in the tephigram. It is readily computed that this lapse rate is equal to about 3/16 of the dry adiabatic lapse rate. A cloud which subsides through its environment will tend to evaporate or to thicken according as the lapse rate in that environment is greater or less than 3/16 times the dry adiabatic.

Of the three cases represented in the diagram, case (2) is the most frequent, case (1) representing an unstable condition, while case (3) includes an inversion below the cloud, which occurs very rarely, if at all.

The above discussion applies to the effects of radiation alone. It cannot be applied to a cloud at an active front. Further, it was assumed that the cloud will gain heat by absorption at its base, and will lose heat on balance by radiation at its upper surface, the rate of loss at the upper surface exceding the rate of gain at the lower surface. This will not necessarily be true if there is an upper sheet of cloud above the one we are considering. In the presence of another sheet of cloud above it, the clearing of the lower cloud will be slowed down, and may be entirely prevented. Should the upper cloud dissipate, the lower cloud may subsequently dissipate.

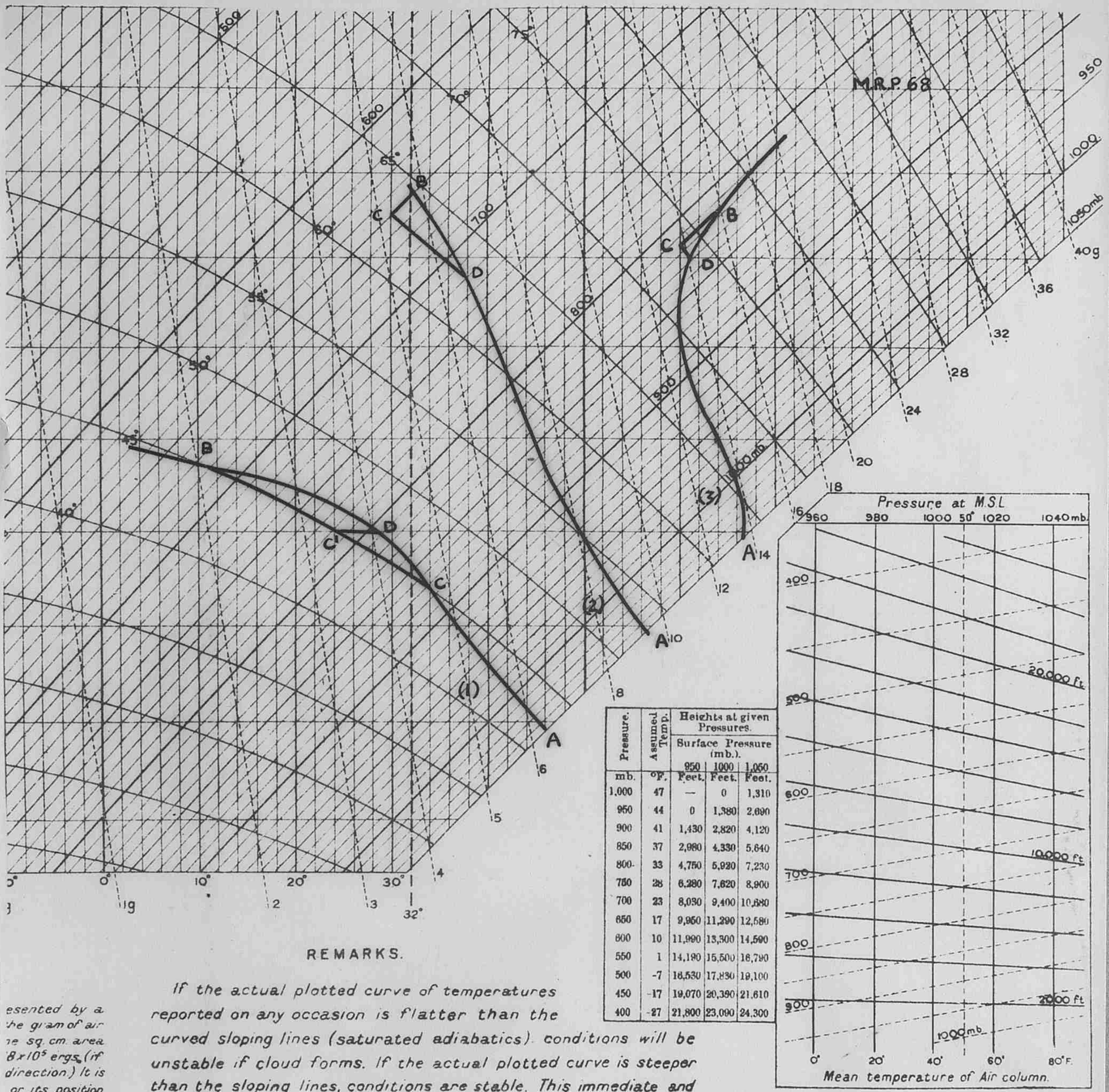
It should be noted that, on the basis of the theory given above, the cloud should become lower before it dissipates. It is not possible to give a time scale for the changes which will occur, but it is clear that the greater the lapse rate below the cloud, the more rapidly will the dissipation be completed. Moreover, the theory can only be applied to the dissipation of the cloud in existence, and it is not to be inferred that the radiational dissipation of an existing sheet of cloud means that other cloud, possibly at totally different levels in the atmosphere, will not form later.

The argument stated in the preceding paragraphs is not applicable to the dissipation of a cloud subsiding *with* its environment. In this case, the representative point of the cloud in the tephigram descends along a saturated adiabatic, while in all cases the cloud becomes thinner, and eventually dissipates if the subsidence is sufficiently long continued. The dissipation, as in the cases already discussed, is produced by the action of two separate factors, (a) the rise of temperature due to increased pressure, and (b), the partial mixing of the cloudy air with the warmer unsaturated air beneath the cloud, produced by the instability flow at the base of the cloud. (A brief discussion of the shear effect in such clouds was given in M.R.P.17.)

References

Durst, C.S., (1933), Q.J.Roy.Met.Soc., LIX, 125

Brunt, D. (1937), ibid, LXII, 276



Pressure.	Assumed Temp.	Heights at given Pressures.		
		Surface Pressure (mb.).		
		950	1000	1,050
mb.	°F.	Feet.	Feet.	Feet.
1,000	47	—	0	1,310
950	44	0	1,380	2,690
900	41	1,430	2,820	4,120
850	37	2,980	4,330	5,640
800-	33	4,750	5,920	7,230
750	28	6,280	7,620	8,900
700	23	8,030	9,400	10,680
650	17	9,950	11,290	12,580
600	10	11,990	13,300	14,590
550	1	14,190	15,500	16,790
500	-7	16,530	17,830	19,100
450	-17	19,070	20,390	21,610
400	-27	21,800	23,090	24,300

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i.e. for 2 sq. cm.

REMARKS.

If the actual plotted curve of temperatures reported on any occasion is flatter than the curved sloping lines (saturated adiabatics) conditions will be unstable if cloud forms. If the actual plotted curve is steeper than the sloping lines, conditions are stable. This immediate and obvious indication of stability or instability is one of the outstanding advantages of the TEPHIGRAM For fuller explanations see Form 2810

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

M.R.P.75
24 February 1943

Further Applications of Wet-Bulb Potential Temperature to Air Mass Analysis.

by F. Tyler, B.Sc.

Hewson has shown the value of the use of wet-bulb potential temperature in air mass analysis. It has further uses as an alternative to the American use of isentropic analysis.

American meteorologists plot specific humidities on an isentropic surface. By plotting wet-bulb potential temperature on an isobaric surface, similar regions of high and low moisture concentration are observed. This is to be expected for:-

(1) Providing sections are taken at similar heights above the surface of the earth, isobaric and isentropic surfaces must give similar humidity distribution. The advantage of using isobaric surfaces is the fact that the readings from aerological ascents are taken on isobaric surfaces, which results in a considerable saving of time in plotting, a fact of great importance in practical forecasting.

(2) Low values of wet-bulb potential temperature will correspond to low values of specific humidity either

(a) because the air is cold and therefore holds little water vapour or (b) because the air is very dry.

The same results may be obtained by plotting actual wet-bulb temperatures on the isobaric surface thus reducing the operation to a matter of seconds, since it is not even necessary to plot the ascents on the tephigram.

Fig.1 shows the distribution of wet-bulb potential temperature on the 850 mb. surface for an anticyclone. It indicates that colder air has been drawn into the circulation by the stronger current further from the centre. This is confirmed by the upper wind trajectories (Fig.2)

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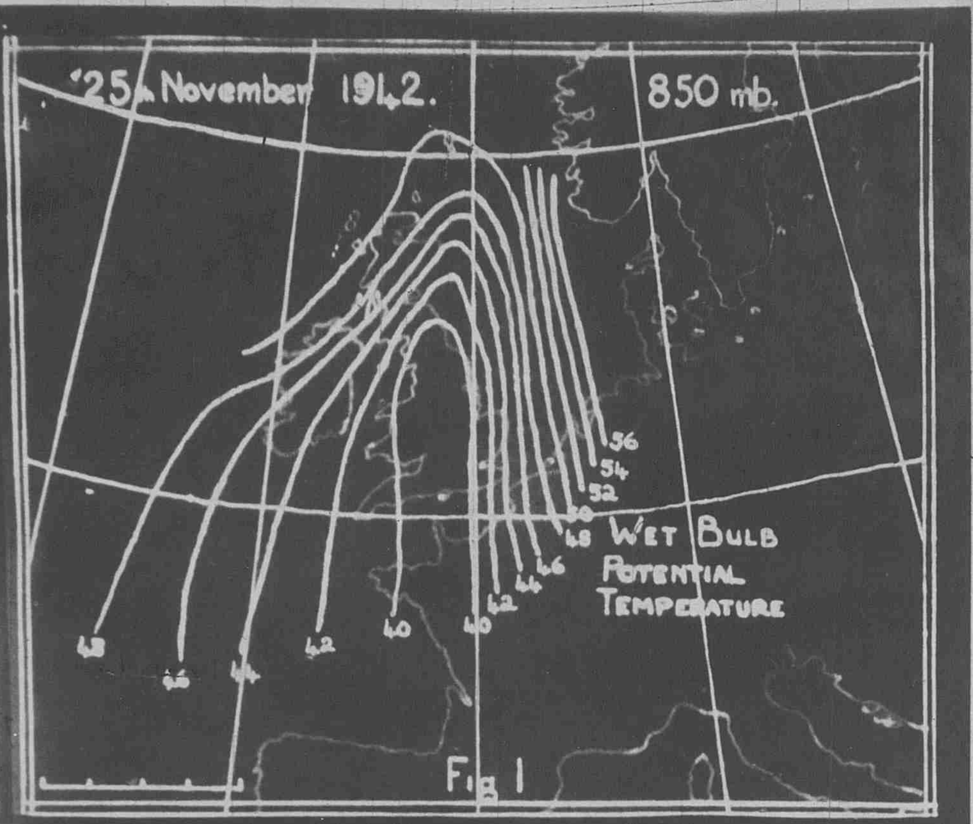
Remarks by D.M.O upon a suggestion by Mr. F. Tyler to plot wet-bulb temperatures on isobaric charts.

The method suggested by Mr. Tyler of plotting the wet bulb potential temperature on the isobaric surface is in current use in the Upper Air Unit at the Central Forecasting Station.

The most useful elements for expressing the humidity in upper air charts are specific humidity, wet bulb temperature or pseudo-wet bulb temperature, potential wet bulb temperature. The most useful element depends to some extent on the type of chart used but on the pressure contour chart the isotherms of wet bulb temperature and potential wet bulb temperature are coincident so that it is a matter of indifference which is plotted from the point of view of distribution. The choice of element to plot depends on what absolute values are required.

A memorandum on the application of humidity in upper air anlysis is in course of preparation and will shortly be issued.


25th November 1942.
850 mb.
56
54
52
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42
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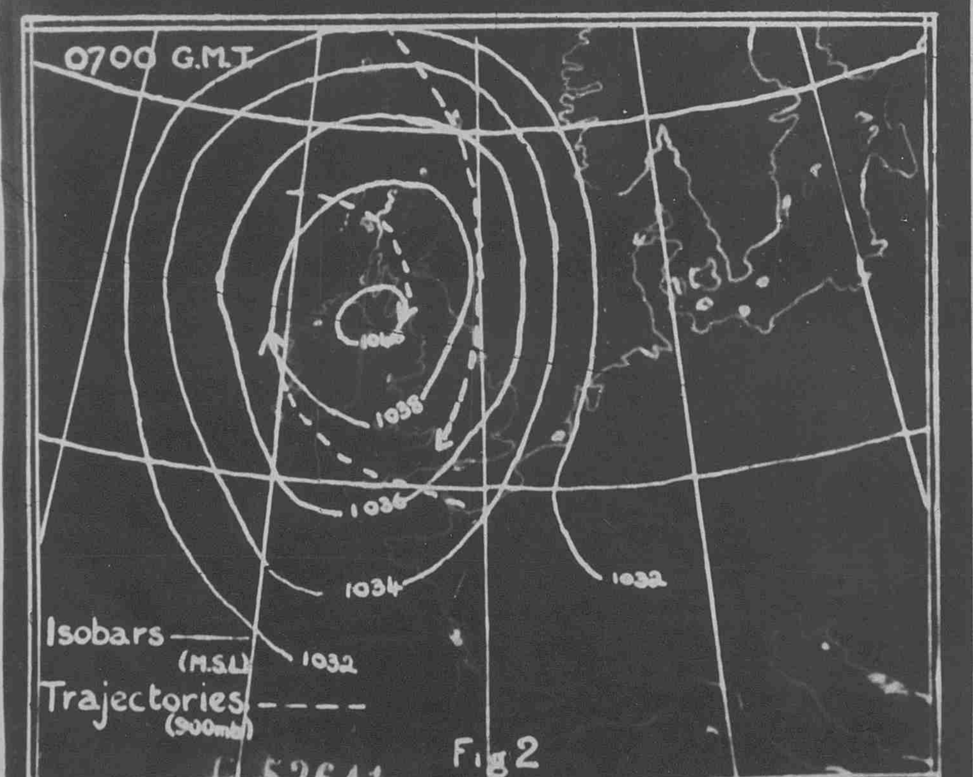
48
Wet Bulb
Potential
Temperature

Fig 1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1032
Isobars ——
(M.S.L.)
Trajectories - - - -
(900mb)

Fig 2

52641

M.R.P. 80
8th March, 1943.

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

CLASSIFICATION OF DAILY WEATHER CHARTS 1930 - 1941: PERSISTENCE AND MOVEMENT OF THE MAIN HIGH PRESSURE SYSTEMS

BY

F. M. JONES, B.A.

CONTENTS

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1. INTRODUCTION

Previous classifications of weather types have been made by Köppen and van Bebber[1], Gold[2] and Newnham[3]. The first and last, and also the present investigation, take account of conditions over the N. Atlantic and Western Europe. Gold's system is based on the detail over the smaller area of the British Isles and France, and several of his types might occur in succession as different phases of one of the types of the present classification.

The development of the Bergen technique, with its improved understanding of the structure of cyclones, has diminished the need for having recourse, in short-period forecasting, to a "precedent book" of previous synoptic situations, but the possible value for longer period forecasting remains. It is reasonable to suppose, however, that a large area must be taken into account in making the classification, and that the types, since there must be a limit to their number, must be chosen on correspondingly broad lines. The types used in this work, which were proposed by Dr. Goldie mainly from a consideration of the results of earlier classifications, and in particular the frequencies tabulated in Mr. Newnham's Geoph. Mem. No.26, are as follows:

TYPE I: Anticyclone in Azores region (and usually also across to Bermuda), i.e. South of 50° N.

IA: without other important high pressure over Europe or N. Atlantic.

IB: with high pressure also in the latitude of Greenland.

IC: with extension towards Iceland.

ID: with extension towards Russia (including extension over British Isles).

TYPE II: The main high pressure system in the latitude of the British Isles (50° - 60° N.), whether centred over the British Isles or W. Europe, or adjacent Atlantic region.

TYPE III: The main high pressure system north of lat. 60°.

TYPE IV: Isolated European anticyclone centred south of 60° N., i.e. southerly or southeasterly type for the British Isles.

TYPE V: A very large anticyclonic area - type occurring seldom except in mid-winter.

The classification is based on the position of the main high pressure systems.

Type IA is the basic or normal type of pressure distribution, and the others can be regarded as variants (more or less wide) produced either by an extension of the sub-tropical High Pressure belt in some direction, or by displacement of the belt. A classification of this sort appears to give a good physical basis for analysis even though as a classification it is not quite exhaustive. In practice about 2.4% of cases could not be fitted into it.

The Northern Hemisphere Synoptic Charts in the British section of the Daily Weather Reports were used for this classification. The daily charts for the period March 1930, when this series commenced, to February, 1941, were classified according to both Newnham's scheme and Goldie's scheme, the results of the latter being set out in Tables 1 and 2a. By using an approximate conversion system, Newnham's tabulated results for the period 1896 to 1910 were converted into the present system, the results being given also in Tables 2b and c.

Again, by using a system of grouping, (which in this case could be only an approximation) Gold's frequency table for the period 1905-18 was converted into a frequency table in terms of Goldie's types, the results being given in Table 3.

2. FREQUENCY OF OCCURRENCE - 1930-41, 1896-1910.

The types of greatest frequency for the year as a whole are IA, II, ID and IB, recurring on an average of 97, 82, 71 and 50 days respectively. Type IA (high pressure in the region of the Azores) predominates notably in summer but is common all through the year; II (Azores high pressure displaced northwards to 50° - 60° N.) is common in all months, but less so in summer; ID (high pressure in Azores with extension towards Russia) figures rather more frequently as a winter type, and IB (high pressure in Azores with high pressure also in latitude of Greenland) is essentially a spring type.

Of the less common types, III (easterly type for the British Isles) is a spring type with a secondary maximum in autumn or early winter, and is seldom found in mid-summer; IC (Azores high pressure with extension towards Iceland) occurs mainly in summer, and IV (southeasterly type) mainly in autumn and winter. V occurs seldom at any time.

Figure 1. shows in greater detail the distribution throughout the year. It was prepared by totalling the frequencies of occurrence of the various types in 5-day periods, January 1st-5th, 6-10th, 11-15th, etc. throughout the 11 years 1930 - 41. Whilst eleven-year totals cannot be expected to give very smooth results, it is of interest to examine this diagram in relation to the "singularities" of weather recognized by Brooks from the study of 52 years of climatological data. The mean dates of Brooks' fair anticyclonic periods have been marked by ⊙ and the mean dates of stormy periods by x. In nearly every case it is seen that the fair periods are in fact associated with more than usual frequency of one or more of the types II and ID which, in general, are fair weather types, and III which can be a fair weather type, and with lessened frequency of IB which is liable to be a stormy type for the British Isles, whilst all the stormy periods are connected, almost exactly, with maxima of IB. The two prominent summer maxima of IB, namely around 8th May and 12th July, do not, however, figure amongst Brooks' stormy periods, nor does the rather sharp maximum of type II at the expense of IA and IB around 26th August figure amongst his fair periods.

Another point which the diagram brings out is the comparative infrequency of type ID - the extension of the Azores High in the direction of Russia - from March to May inclusive, i.e. in the period of the year when the land surface is heating up. The building up of the Azores High in summer is also marked both in its IB form and more particularly in the form of type IA in July and August, with at the same time the virtual exclusion of types IV and III respectively.

As mentioned above (page 1), the more localised types of Gold's classification occur as different phases of the types of the present broad classification. Although it is not possible to define exactly the connection between the two sets of types, an approximate grouping has been worked out.

Some idea of the frequency with which Goldie's types should occur can be seen from Table 3 which has been compiled from Table III of Gold's Geoph. Mem. No.16 according to the conversion system indicated in the first two columns.

It will be seen that the combination IA, IB is the most frequent type in every month of the year except May and October, which are in some ways transitional months. The frequency is highest in July and August, and gives a total of 136 days for the year. For the periods covered in the present memoir, the combination IA, IB has its highest also in July and August, and the total for the year is 147. The broad resemblance of the rest of the table to table 2c will be noted.

3. DURATION OF RUNS; SUCCESSION; RELATIVE PERSISTENCY OF TYPES.

Table 4 shows the frequency of runs of given durations, and the mean duration of each type month by month, for the period March, 1930 to February, 1941. 50 per cent of the days on which type IA occurred were in runs of more than 4 days; 44 per cent of IB, 0 per cent of IC, 28 per cent of ID, 57 per cent of II, 57 per cent of III, 35 per cent of IV and 0 per cent of V. Mean durations of 4 or more days are shown by IA in August, IB in May, II in December, January, February, April and May, III in January, February, March, May and June. Though type II occurs very frequently in September, its mean duration does not rise in that month.

Table 5[*] is an analysis of the succession of types, the figures being percentages of the total number of occurrences. As might be expected, except in the case of the rather infrequently occurring types, the tendency is for any given type to be followed most often by itself.

TABLE 6 - Persistency Index 1930 - 41

	IA	IB	IC	ID	II	III	IV	V
WINTER	.59	.54	.18	.50	.73	.69	.39	0
SPRING	.57	.59	.29	.44	.73	.70	.48	0
SUMMER	.45	.56	.23	.49	.65	.68	0	0
AUTUMN	.59	.47	.28	.49	.64	.38	.60	0

Table 6 shows the seasonal value of the "persistency index"[ø] for each type. Type II is the most persistent in all seasons, with the exception of type III in summer: i.e., a high pressure system centred between 50° and 60° N., tending to be displaced northwards during the summer months. Types II in all seasons, III except in autumn, and IA in all seasons are persistent, ID rather less, whilst IC and V are non-persistent as well as infrequent. Taking all types together, the greatest duration tends to be in January and June, with minima of duration April and September.

Next to the tendency for a type to persist, there is a distinct inclination for types IB, IC and ID to relapse back to IA, i.e. the most likely change is for the subsidiary high pressure in the latitude of Greenland or the extensions from the Azores towards Iceland or from the Azores towards Russia to die out and leave only the main Azores High; alternatively, IC may develop into ID or II. Type III on changing is followed most often by type II, but sometimes changes to IB. In so far as it shows any definite tendency, type IV most frequently develops into ID. The above deductions are drawn purely from the statistics of tables 5 and 6, and take no account of the duration of the type. Cases where a type has persisted for a time are dealt with below in greater detail.

Consideration of runs of five or more days of a given type leads to table 7.

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[*] See also table on page 25 of Köppen and van Bebber[1], Archiv der Deutschen Seewarte, No.4, 1894.

[ø] "Persistency Index". $P = \frac{y - x}{100 - x}$, where x = the total number of days that the type has occurred in any season, as a per cent of the total number of days in that season; y = per cent number of times on which a type is followed by itself. (see table 5)

/TABLE 7

TABLE 7 - Percentage number of times a run of 5 or more days of a type has been preceded or followed by any other type. (Types IC and V did not show any runs of as much as 5 days).

BEFORE									AFTER							
IA	IB	IC	ID	II	III	IV	V		IA	IB	IC	ID	II	III	IV	V
-	23	3	40	25	0	9	0	IA	-	27	6	35	25	3	3	0
37	-	10	19	21	9	4	0	IB	36	-	9	26	16	8	6	0
29	28	2	-	22	5	14	0	ID	37	17	3	-	27	5	11	0
26	9	4	39	-	18	2	1	II	28	19	9	18	-	18	9	0
9	37	6	9	19	-	20	0	III	6	44	0	20	20	-	9	0
15	20	0	30	25	10	-	0	IV	20	15	0	55	10	0	-	0

This shows that IA, IB, ID or II are most often both preceded and followed by one of this same group of types, whilst III, IV and IB similarly can be grouped as regards sequence. Cases of persistence for 5, 10 days or longer are dealt with in greater detail below.

4. INCIDENCE AND PERSISTENCE OF HIGH PRESSURE IN CERTAIN AREAS

Type I - taken as the total of IA, IB, IC, ID - occurs in the mean per annum for 1930-41 on 228 days; i.e., the Azores High is present, with or without extensions, on nearly two-thirds of the days of the year (see table 8). When this type is specially frequent in any year it is almost entirely at the expense of type III and vice versa. On the other hand, all months, when there has been a notable deficiency of type I, have been characterized by a somewhat corresponding excess of type II so that presumably a deficiency of type I means merely a displacement of the High in latitude.

Table 8 shows the frequency of runs of type I of different durations and the mean duration; it is clear that the degree of persistence of high pressure in the Azores region far exceeds that of high pressure in any other region. The frequency is highest in July and August (29 and 26 days) and lowest in December and September (14 and 15 days). The runs vary in mean duration from 21 days in July to about $5\frac{1}{2}$ days in December. The longest spell recorded in December was 14 days. Long spells have been less common in December and around the equinoxes. At other times of the year spells exceeding 30 days have occurred.

TABLE 8 - Number of cases of runs of Type I.

Days	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Year
1- 5	11	9	9	11	7	9	2	5	14	21	21	17	
6-10	4	2	9	4	6	3	2	4	9	5	2	7	
11-15	7	1	1	3	4	5	2	4	2		5	3	
16-20	1	2	2	3		1	2	1	2	1	1		
21-25		1		1		3	2			1			
26-30	1				2		1				2		
above 30		2	1		2		4	4		2	1		
No. of runs	24	17	22	22	21	21	15	18	27	30	32	27	
No. of days	194	187	174	169	232	188	317	284	166	216	233	150	2510
Mean duration	8.1	11.1	7.9	7.7	11.0	9.0	21.1	15.8	6.4	7.2	7.3	5.6	
Spells of 11 or more days	9	6	4	7	8	9	11	9	4	4	9	3	

For types II and III, runs of over 10 days are much more common in the first six months than in the second six months of the year (8 against 3 and 4 against 0). Long runs of type II attain their maximum frequency from January to May, and type III has a similar persistence only from January to March. Apparently late winter or late winter and spring conditions are most favourable to this development in which anticyclones in the latitude 50° - 60° or 60° - 70° replace or overshadow the Azores High. This points to cold or relatively cold surface conditions as a requirement.

Of the occasions where type II persisted for 5 days or more, it was both preceded and succeeded by type I on 55% of these cases, preceded by I and followed by III on 16%, preceded by III and followed by I on 11%, and both preceded and followed by III on 4%; that is, the type II mostly developed from type I and mostly relapsed into type I, but in some cases it developed from or into type III.

Consistently with this, it is found that Type III was in less than 20% of cases a development from type II. In seven cases out of nine in summer and in more than half the cases in winter, when it lasted for 5 days or more, it was a development from I (usually IB - the type where in addition to high pressure in the Azores region there is high also in the latitude of Greenland). Similarly type III in 8 cases out of 9 in summer and in 60% of the cases in winter changed back into type I. Type III, in fact, is mostly a development of type IB, but in winter there is alternatively a chance of its developing from or into type II.

Where type II was more persistent and lasted for 10 days or more in 8 of the 9 summer cases and in 6 of the 10 winter cases it developed from the Azores High or arose actually as a northward displacement of the Azores high. In 3 of the remaining 4 winter cases and the 1 summer case a high from Iceland or Greenland drifted southward.

In the persistent cases of type II in the summer six months the Highs are nearly always centred to the W. or N.W. of the British Isles or over the British Isles; relatively insignificant changes of pressure are sufficient to change the position of the centre within the area in question. In the winter the highs in about half the cases are in the same position; in other cases they may be centred over the Channel or to S.W. of Ireland, or farther W. on the Atlantic.

When type III is persistent for 10 days or more it is very often the case that the highs lie over Greenland or Scandinavia or drift eastward from Greenland, but there seems to be a liability to fairly rapid displacement.

In the matter of upper air temperature, since the Highs in type II were in most cases centred to westward of the British Isles, the only temperature observations available are those made in the British Isles, i.e., in general in the northerly current. In summer the average temperature for the cases of duration 10 days or longer is slightly below normal at 900 mb. and about one and a half degrees F. above the normal for the month from the 800 mb. to the 450 mb. level. This means that the air is on the whole rather cold for anticyclones. Roughly, the same is true in winter, the mean for all the cases gives a temperature one and a half degrees below the monthly normal at 900 mb. and about $2\frac{1}{2}°$ above the monthly normal from 800 mb. to 450 mb.

As regards individual cases, some have been cold and some warm without any apparent relationship between temperature and duration. The two highs, which in winter drifted southward from Iceland, were distinctly cold, especially in the lower layers; one which came from Greenland was cold in the lower layers and warm at higher levels. On the other hand, some of the highs which have appeared to spread northward from the Azores were cold at all levels up to the 450 mb. level. In such cases the apparent drift of the High northward from the Azores was connected with the spread of a strong cold current from the north over the British Isles. There was no transport of air from S. to N., at least below the 450 mb. level; the main transport was in the opposite direction.

5. RECURRENCE OF TYPES AT REGULAR INTERVALS

Related to the question of persistence of high pressure systems is the question of possible recurrence at more or less definite intervals.

Though the normal circulation of the atmosphere of the Northern Hemisphere is much less regular than that of the Southern Hemisphere, the movement of the Northern air cap at the 4 km. level, according to A.W. Lee, 4, can be regarded as approximately made up of a solid rotation once in 19 days in winter and once in 38 days in summer, with certain superposed systems. The chief of the superposed systems are, a belt of high pressure between latitudes 50° and 65° extending from the British Isles across Europe and Siberia; highs over N.W. America and over Algeria; lows to the South of Greenland and over the Eastern Pacific. At 6 and 8 km. in summer the mean rotation times are 33 and 23 days respectively. As regards the Southern Hemisphere, E. Kidson, 5, concluded that if the mean velocity and distribution of anticyclones over the Australasian region can be taken as representative, an anticyclone would on the average complete a circuit of the globe in 42 days; this is for a lower latitude, namely about 35° S.

In view of these results, it seemed worth examining the question whether occasions of type II or type III, that is, the main high pressure system in latitude 50° to 60° or northward of 60°, rather than in the usual latitude of the Azores, show any tendency to recur after periods of the order mentioned above. Clearly, if there were something in the high atmosphere which favoured the occurrence of higher pressure or lower pressure than usual, and if that something persisted long enough, some recurrence tendency with a period of this order is to be expected. On the other hand, the low degree of persistence commonly shown by pressure systems other than the subtropical high and the infrequency of any long continued regular west to east travel in the northern hemisphere make it less likely that recurrences would readily be discovered at intervals equal to the rotation period of the circumpolar vortex.

The results of tests applied to check the recurrence of type II are set out in figures 2a to e. Figure 2a shows the number of occurrences of type II at intervals of 1, 2, 3, ... etc. days after an occurrence of type II in spring. Figures 2b, 2c, 2d show similar data for summer, autumn and winter respectively.

Figure 2e shows the recurrences of cases of Atlantic depressions with isobars for pressure under 970 mb.

Taking these diagrams in turn it is seen that Figure b (Summer) actually shows a maximum recurrence of type II at about 38 days after the first occurrence and another at about 66 days. The first maximum amounts to only about 1/3rd of the initial occurrences and it is not repeated in any of the other curves. Curiously enough, however, the maximum of recurrence at 66 days on the Summer curve is in accord with the maximum on the winter curve, whilst any tendency which the autumn curve shows towards a maximum is near the same place, namely at about 57 days. The spring curve shows no evidence of recurrence at definite intervals.

Curves (not reproduced here) were prepared also for cases of initial spells of 4 or more days of either type II or type III. These were relatively flat, though those for autumn and winter gave a feeble maximum at about 28-30 days and the summer curve was not inconsistent with a maximum in this region.

Curve e, relating to recurrence of very low pressures, shows nothing beyond a gradual falling off of frequency with time as measured from an occurrence of pressure below 970 mb.

The final conclusion is that these data yield no evidence sufficient to confirm a tendency for displacements in latitude of the main high pressure system to recur with any regularity at definite intervals, but on

/the other

the other hand they are not inconsistent with the possibility that peculiarities of the upper atmospheric circulation do at times persist long enough to recur at intervals of the order of the rotation periods computed by Lee, or Kidson.

That the evidence of recurrence is seen chiefly in the summer and winter curves adds to the probable significance since the principal "symmetry points", so far recognised in regard to pressure distribution, are those near the end or beginning of the year and in the month of June.

6. VARIATION OF THE AZORES HIGH AND THE ICELANDIC LOW.

CIRCULATION INDEX.

The classification of types used in this work turns largely on the position and intensity of the Azores High and its displacements. Table 9 contains the mean latitude and maximum pressure in the Azores High, the mean latitude and minimum pressure in the Icelandic Low and the mean gradient in each month as estimated from the Monthly Charts of Pressure of the Western North Atlantic (M.O.M.394) and the Charts of Home Waters (M.O.M.447). The pressure values on these charts are means for the period 1921-38. There are two almost distinct levels of flow with rapid transition about March and October, the mean flow West-East in the months November to February inclusive being about 70% above the mean level in the months March to October inclusive.

In order to follow short-period changes in the circulation, a tabulation was made for each day of the highest pressure within the area 10° W. - 70° W. and S. of 50° N. and the lowest within the area 20° E.-60° W. and N. of 50° N. On most days this means the highest pressure in the Azores High and the lowest pressure in the Icelandic Low respectively. The difference between these pressures is a measure of the total (approximately) West-East flow of air across the Eastern Atlantic towards Europe.

It is an index somewhat similar to but not quite the same thing as Rossby's* zonal circulation index.[6] It attempts to assess the total flux of a great air current which is practically always in existence, but is somewhat variable in width as well as in position. Because of its inertia it may be expected to show a certain degree of persistence. The mean day-to-day fluctuation is in this case found to be about 16% and it is seldom that twice this rate of change is exceeded. When five-day means are considered the mean rate of change per 24 hours ranges from some 2% in quiet periods to about 5% in very disturbed periods.

The mean and extreme values of the 5-day means for the nine years September, 1930 - August, 1939 inclusive are shown in Fig.3. The mean dates of Brooks' fair and disturbed periods are marked by ⊙ and x respectively.

On account of scarcity of data from the Atlantic after the onset of the war it was considered desirable to omit the data from September, 1939, in computing these means and extremes.

Apart from a certain persistence, there is at first sight no evidence of regularity in this flow other than in the form of a strong seasonal variation similar to that shown in Table 9. When, however, an examination is made for recurrences of high or low values in the flow, some interesting features are noticed. The mean flow index during the 6 winter months is 47 mb. All cases were marked where the three-day mean reached or exceeded 68 mb. or where there was an individual value of 72 or more.

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* Rossby's zonal circulation index is computed as the mean difference of pressure between lat. 35 and lat. 55 and is thus a measure (approximately) of mean West-East component of wind between these fixed latitudes.

All these cases were in the winter six months. Then after each such case the number of recurrences of similar maxima at 1, 2, 3 .... 72 days after was counted. These were then grouped in three-day totals centred at 3, 6, 9, 12, 15, etc. days after.

Minima of flow - cases of the three-day mean being 33 or less - were dealt with similarly. The resulting curves and their mean are plotted in Fig.4(a). The result suggests a tendency to recurrence of similar conditions of flow after intervals of 6 and 18 to 24 days fairly definitely, probably about 30 to 33 and possibly also around 66 to 72 days, the last mentioned being similar to the long period noted in both the winter and summer recurrences of weather type II (Section 5).

A similar examination of the data of the summer six months led to the curves in Fig.4(b). Here there is evidence of recurrences after intervals of 6 and about 24-30 days and possibly about 72 days.

The existence of periods of these orders is in accord with the results of Weickmann, Wadsworth and others who have carried out harmonic analysis of pressure charts.

The prominence of the interval of 6 days (or to be exact, something - possibly not fixed - lying between 5 and 7 days) suggests that it is the fundamental period; very possibly the recurrences at longer intervals, which are approximately integral multiples of the 6-day period, are brought about by the 6-day recurrence tendency.

Rossby[6] suggests that for the average West-East wind velocities prevailing in middle latitudes, the wave-length should be of the order of 3000 miles, more in winter, less in summer. The above 6-day period in the circulation index, taken in combination with the most common speeds of depressions in winter and summer, namely 25 and 20 m.p.h. respectively (see Section 10) points to common winter and summer wave-lengths of the order of 3600 and 2880 miles, which is in good agreement with Rossby's figure.

(Tabulations of daily values and of five-day means of the circulation index for the period March, 1930 - February, 1941, are available and may be consulted on application to the Director of the Meteorological Office.)

7(a). INTENSE DEVELOPMENT OF ATLANTIC HIGH

In 84 cases in the 11-year period there were isobars of 1040 millibars or higher in the Azores to Bermuda region or over the Atlantic south of latitude 50° north. The distribution of these among the various types is shown below together with the percentage which those cases represent in relation to the total number of occurrences of the various types.

TABLE 10 - Distribution of occurrences of 1040 mb. or higher in Azores.

Type	No. of cases of Atlantic pressure 1040 millibars or over.	Relation to total number of occurrences of the type.	No. of cases of Atlantic pressure 1044 millibars or over.
IA	24	2.8%	3
IB	2	.3	-
IC	4	3.0	-
ID	13	1.7	4
II	33	3.5	5
III	6	1.6	1
IV	1	-	-
V	1	-	-

/Thus

Thus pressures of 1040 millibars or over on the Atlantic occur principally with type II (high pressure centred in the latitude of the British Isles) or with type IA, or the relatively infrequent type IC.

The rekative infrequency of these high pressures over the ocean (south of latitude 50°), however, will be noted. They occur almost entirely between November and April. The years which had the most frequent occurrence of very low pressures are for the most part the same as those which most frequently had very high pressures.

7(b). INTENSE DEVELOPMENT OF THE ICELANDIC LOW.

In 266 cases in the eleven-year period, there were isobars of 968 millibars or lower in the Icelandic region and in 36 cases isobars of 958 millibars or lower. The distribution of these amongst the various types is given in table 11, together with the percentage which these cases represent in relation to the total number of occurrences of the various types.

TABLE 11 - Distribution of occurrences of 968 mb. or lower in Icelandic region.

Type	No. of cases of Icelandic pressure below 970 millibars.	Relation to total No. of occurrences of the Type.	No. of cases of Icelandic pressure below 960 millibars.
IA	80	9.1%	4
IB	16	2.2	1
IC	3	2.3	1
ID	81	10.4	17
II	41	4.3	5
III	11	2.9	2
IV	34	18.8	6

Clearly the maximum development of the Icelandic Low is with types IV and ID, and next to these comes IA.

None of the cases occurred in the months May to August inclusive; 73 occurred in January, 27 of these being in the week January 12th-18th.

8. CHARACTERISTICS OF DIFFERENT YEARS AND SEASONS.

(a) Years. Runs of type II have occurred with high frequency in every year, but the frequency was rather higher from 1933 to 1936 than in other years between 1930 and 1941. Type IB also occurred with more than usual frequency in 1935 and 1936.

The incidence of runs of IA has varied greatly; 1931, 1936, 1937 and 1939 were very deficient in runs of this type, whilst 1940 had an unusually large number.

(b) Seasons. The average frequency of occurrence of each type in each season is set out in Table 12 and the departures from these averages are given for the principal types in table 13.

The Azores High was more than usually predominant in the following seasons:-

Spring 1930 - type IA chiefly, at the cost of type II.
Winter 1930-31 ID II, III, IV.
Autumn 1932 - IA II and IV.
Autumn 1940 - IA IV.
Winter 1940-41 IA III and IV.

In the autumn and winter of 1935-36, there was an unusual preponderance of IB at the cost of IA and ID; this was a stormy unsettled period.

The Azores High was deficient in the following seasons, usually, as will be noted, with an excess of type II.

Autumn 1931 and Winter 1931-32, with great excess of type II.
Autumn 1933 and Winter 1933-34, with excess of type II.
Spring 1935 with great excess of type II.
Spring, Summer and Autumn 1937 with excess of either III or II.
Spring 1938 with great excess of type II.

Type III occurred with remarkable frequency in the winter of 1939-40; for a year afterwards and especially (as noted above) in the autumn of 1940 and winter of 1940-41 there was unusual preponderance of IA.

Type IV occurred with more than its usual frequency - at the cost of type II - in the autumns of 1935 and 1938 and more notably the winter of 1938-39.

(c) Related Occurrences. Whilst it is not easy to see any definite relationship between one season and another, there are a few correlations.

Relatively high frequency of type II in spring tends to be followed by infrequency in the autumn and vice versa. In the 11 years 1930-41, the correlation between number of occurrences of type II in spring and number in autumn is -.68 with a probable error of .11; between spring and summer plus autumn the coefficient is -.77 with a probable error of .08. The nature of the connection is probably through ice distribution in the Greenland-Iceland region (see Appendix). Another kind of sequence, taking account of temperature conditions over the British Isles is noted below.

(d) Abnormally Warm Months. In general, months which are warmer than normal over the British Isles are associated with an excess of types IA and ID and a somewhat corresponding deficiency of type II. These type features of the warm month itself are, however, less outstanding than the conditions preceding and succeeding the warm month. The table below shows the departure from normal temperature over the British Isles in the ten warmest months of the eleven-year period and also the relative frequency of occurrence of type II as compared with normal in the months in question and in the preceding and succeeding months.

TABLE 14 - Abnormally warm months related to Frequency of Type II

	Diff. of Temp. from normal °F.	Frequency of Type II in relation to normal		
		Preceding Month	Warm Month	Succeeding Month
Jan. 1932	+4.9°	+12	+2	+17
Dec. 1934	+5.5	+8	-8	+10
Mar. 1938	+6.1	+9	+2	+21
Nov. 1938	+4.7	-3	-7	-8
June 1940	+4.3	+1	-5	0
Mar. 1933	+3.7	-1	-6	+3
July 1933	+3.5	-4	-1	+2
Aug. 1933	+3.7	-1	+2	+3
Sept. 1933	+3.3	+2	+3	+5
July 1934	+3.8	+5	+3	-4
Mean percentage departure from normal		+36%	-20%	+60%

Since a warm month is as a rule associated with a specially well developed Azores High, the result shown in table 14 - that the warm month is followed mostly by a month with an excess of type II - is in accord with Brooks' conclusion[7] from pressure anomaly maps, namely that there is about a 4 to 1 chance that a high positive anomaly in one month over the Azores will travel northeastward and appear as a high positive anomaly over the British Isles in the following month.

The months which followed the three most abnormally warm months had respectively 25, 17 and 30 days of type II.

(e) Abnormally cold months. Abnormally cold months are associated mostly with a deficiency of type ID and an excess of either II or III. The conditions preceding and succeeding them are less uniform. In the 11-year period under review there have been four months with a temperature deficiency exceeding 3°F. over the British Isles; the conditions were as in table 15.

TABLE 15 - Abnormally cold months related to sequence of types.

Month	Diff. of temp. from normal	Weather types in relation to normal		
		Preceding Month	Cold Month	Succeeding Month
Jan. 1940	-7.7°F.	Nothing of note	ID deficient; excess of III	ID and II deficient; excess of III
Jan. 1941	-6.2°F.	ID deficient; excess of IA, IB	ID deficient; excess of II, IA, IB	ID deficient; excess of IA, IB
Mar. 1937	-3.7°F.	Excess of III and IV	ID deficient; excess of III, IV	Excess of III, IV
Dec. 1933	-3.1°F.	ID deficient; excess of II	ID deficient; excess of II	Excess of ID

9. WEATHER ASSOCIATED WITH VARIOUS TYPES.

Using the Observatories' Year Book, the weather experienced at Kew Observatory and at Valentia on occasions when the same type lasted for three or more days has been tabulated and analysed. The results are summarized in tables 16(a)-(g) and 17.

These tables show that for the London area, type II is on the whole the driest, with 75% of absolutely rainless days, and with average daily falls ranging from 0.1 mm. in February to about 1 mm. in June, August and September. The larger amounts are due in part to thunderstorms, but thunderstorms with this type are relatively less frequent than with other types. At Kew 89% and at Valentia 83% of type II days have less than 1 mm. of rain.

With type II, however, Valentia shows a remarkable peak in August in regard to both frequency and amount of rainfall. This is not shown by Kew; Valentia shows a somewhat similar peak for type IA in September in amount of rainfall only. April on the other hand is conspicuously dry and sunny at Valentia with type ID.

Next to type II for dryness comes ID, except possibly in Autumn. Type IB is considerably affected by thunder in Summer and has a higher frequency of thunder than other types; from October to December it gives, however, a low average daily rainfall, and this, together with the dependence of its summer rainfall on thunder, marks it as having conditions in which instability contributes considerably to the rainfall.

Types III and IV are the wettest types in winter at both Kew and Valentia, and have the highest percentage of days with over 5 mm. of rain.

The average daily amounts of rainfall with type III at Valentia in November, December and January are remarkable, some 12 - 14 mm., as against averages of the order of 1 - 3 mm. from March to October inclusive. The frequency at Valentia in winter is also 100%. Kew shows relatively a small annual variation in mean daily rainfall with type III, the values for individual months being of the order of 1 - 3 mm.

Types II and ID are on the whole the sunniest, except in winter when IB and IV stand relatively high.

For the year as a whole, the percentages of days which had 80% or more of the possible duration of sunshine are as follows:-

Type	IA	IB	ID	II	III	IV
Kew	3	3	6	9	1	2
Valentia	1	4	4	16	5	4

In summer months at Kew some 23% of type II days and in June some 21% of ID days have 80% of the possible duration of bright sunshine, or over.

At Valentia 24% of type II days in both spring and summer have 80% or more of possible duration of bright sunshine.

Though Valentia has a higher rainfall than Kew and a higher number of days of rain in winter, the frequency of sunless days is nevertheless lower than at Kew in winter with types ID, II and III. Presumably this is explained by the tendency to fog in S.E. England with these types. With IA the position is reversed.

Type III, particularly in January, February and March, and II in February are the ones most likely to give snow, but IV also has a high liability to snow at Valentia in January.

Type III in August, ID in October, IB in June, August and September have a relatively high liability to thunder at Kew. At Valentia, on the other hand, as on the west coast generally, the liability to thunder is mainly in winter and types IV, III and ID chiefly contribute. In the case of fog the position is completely reversed. At Kew the winter months, with the quiet types II and III, have a high incidence of fog, the westerly types IA, IB, ID being less unfavourable. At Valentia it is only in summer, and then mainly with ID and II (and with III in April) that noticeable amounts of fog occur.

10. TRACKS AND MOVEMENTS OF DEPRESSIONS

(a) Depression tracks in the various types.

The tracks of depressions drawn in the maps in the Monthly Weather Report have been examined in relation to the types with which they were associated. For the most part there is great complexity, but in some cases definite tendencies can be noted. Only a few of the classified tracks are reproduced here (Figs. 5(a) to (h)).

TYPE IA. With an anticyclone in the Azores region and no other important high pressure, it is to be expected that paths of depressions will be somewhat widely dispersed.

/In winter

In winter, Fig.5(c), the tracks cover almost the whole area of the map, but are not spaced with equal density. There are congestions of tracks along the Norwegian coast, between Scotland and Faroe, in the southern part of the North Sea between the longitudes of 0° and 5° E., and at the western end of the English Channel, whilst the land areas of Scotland and Ireland seem to be rather avoided. In Autumn, Fig.5(b), there is less bias in favour of particular areas; the southern parts of the North Sea, S.E. England and France, except the north and west coast, are, however, crossed by scarcely any tracks. In August, Fig.5(g), eastern France and the greater part of England are avoided; and in June the whole area of Ireland, England, the southern North Sea and France is practically clear of tracks, but this is not the case in July which has tracks almost everywhere and running in almost every direction. In Spring there is again a very wide dispersal of tracks.

TYPE IB, IC, ID. The depression tracks for these types have not been drawn in detail, but a statistical summary of directions (from which depressions move) shows the following differences, as compared with IA.

IB - In summer and autumn tracks from S.W. increase compared with IA, whilst throughout the year generally appreciable numbers of tracks from directions between S. and E. are found. In summer tracks are mainly concentrated between the Faroes and the English Channel. In winter they are dispersed over a wider area, from lat. 70-45, but again are mainly from S.W., with some from between S. and E.

IC - The number of cases in winter and autumn is too small for deductions to be justified, but in spring and autumn the chief difference from IA is that appreciable numbers of tracks are found from directions between S.E. and N. (through east).

ID - The most noticeable feature - and it is in accord with the extension of high pressure from the Azores towards Russia and Siberia - is a notable increase in all seasons in the number of tracks from S.W., and a decrease in those from either S. or N.W. In spring and summer there are also a few tracks from between S.E. and N.E.

TYPE II. In Section 4, it has been pointed out that in type II the anticyclone, in fact, lies very frequently over or to westward or northwestward of the British Isles. It is, therefore, in accord with expectation that the tracks tend to surround the British Isles on the north, east and south sides. The classified paths are shown in figures 5(e) to (h). In all four seasons the main track is pushed northward to the latitude of Iceland. In autumn and winter there are definite tracks along the English Channel and off the coast of the Low Countries. In summer the depression centres tend rather to cross the south of England, and thence towards Denmark. Particularly in spring, but also to some extent in winter and less noticeably in summer and autumn, there is a tendency for depressions to move from N.W. to S.E. across the North Sea or Scandinavia.

It is to be noted that even to southward of the main high pressure system most of the depressions move from westerly points, sometimes with a component from north, and relatively few move from easterly points.

TYPE III. Since type III occurs seldom in summer (see table 2), there were insufficient cases from which to make any deductions.

The winter tracks, Fig.5(d), show a considerable number of depressions from W. or S'westerly directions, with two main groups:-

(i) across S. Ireland and S. England, dividing over the North Sea to go N. and S. of Denmark, and

(ii) from S.W. across the west coast of France.

/A-

A third group from N. along the Norwegian coast is associated with the cases where the anticyclone of type III is to the eastwards of 5° E.

TYPE IV. Tracks were not drawn in detail. In autumn 53% of the depressions noted came from between S. and W.S.W. In winter 79% came from between S.W. and W.N.W. The type was absent in summer and there were but few occurrences in spring.

(b) Speed and direction of travel of depressions.

Table 18 gives the arithmetic means of speeds of travel of depressions in the various types.

Tables 19 and 20 summarize the frequency with which different directions and speeds occur in the different seasons.

Table 21 shows the vector means of depression tracks for each type and each season.

In the tables the speeds are given in terms of the distance in miles travelling by the centre in 12 hours, usually the 12 hours from 1 h. to 13 h. or 13 h. to 1 h., but sometimes calculated from the distance travelled between other pairs of synoptic hours.

In every season and with every type there is a great range of speed, but as a rule 50 to 60 % of depressions have a speed between 200 and 400 miles per 12 hours. The arithmetic means in table 18 range from 22 to 37 m.p.h., the mean of all being 26 m.p.h. In summer and spring the highest frequencies are about 20 m.p.h. and in winter and autumn about 25 m.p.h. Taken irrespective of direction, the depressions in types ID and II have the greatest speed, in all seasons except summer. They have also the greatest seasonal range of speed. A minimum of speed is shown by IC in all seasons except in summer.

Despite the considerable range of speed in every season, the arithmetic means, as shown in table 18, do not vary very greatly with season, showing a range as between summer and winter of rather under 20% of the annual mean. This range is in agreement with figures given by Hann-Suring[8] for the West Atlantic and West Europe, though their seasonal mean values are slightly lower.

Palmën[9] gives a formula for the speed of depressions,

$$V = \alpha + \beta v + \gamma \sqrt{\Delta T}$$

where α, β, γ are constants varying, however, with the type and condition of the depression, v is the air speed in the warm sector and ΔT is the temperature difference at the surface of discontinuity. According to values given by Goldie[10], the monthly mean temperature differences between tropical and polar air range from about 5°C. in summer to 10°C. in winter or spring. Palmën's formula, with a value of 2.6 to 3.0 for γ, indicates a seasonal range in mean speed on this account of the order of 8 m.p.h. Since the term βv would also add a little to the seasonal range, it appears that the formula would give a higher seasonal range than that indicated by the tables above. For 245 depressions Palmën's mean speed is 26 m.p.h., exactly the same as in the table above for 1876 cases.

The vector means, as shown in table 21 and figures 6 and 7, vary more considerably in speed, from 30-50% in types IA, IB, IC and ID, though IA and ID, which as might be expected, show the highest vector mean speeds, have almost a constant direction. In types II and III the seasonal mean speed is almost constant.

In no case except type III in summer is there a seasonal vector mean from an easterly point, and the total number of instances (6) in this case is too small to have much significance.

/APPENDIX.

APPENDIX

Relationships between weather types and ice conditions in Barentz Sea and Greenland-Iceland region.

by

Miss M. Hewitt, B.Sc., F.G.S.

The approximate ice limits were charted by the Marine Branch of the Meteorological Office for the months April to August from 1930 to 1938 inclusive. The criterion of ice in the Barentz Sea was taken as the mean of the number of degrees south of latitude 70° N. of the points where the ice limit crossed the longitudes of 20°, 30°, 40° E. respectively. For the Greenland-Iceland region the criterion was the mean of the number of degrees south of latitude 80° N. at which the ice limits cut the longitudes of 20°, 30° and 40° W. respectively. The differences of these figures from the means for each month for the whole period are given in table I.

TABLE I - Differences from monthly mean in the ice in:-

	1. Barentz Sea					2. Greenland-Iceland Region				
	Apr.	May	June	July	Aug.	Apr.	May	June	July	Aug.
1930	0.0	+0.7	+0.2	-0.8	-2.3	-0.2	-0.5	-0.2	-0.7	-0.5
1931	-0.3	-0.4	+0.6	+1.1	-2.0	-0.4	-1.0	-0.2	+0.3	+0.5
1932	+0.9	+1.1	+1.6	+1.2	+0.4	0.0	0.0	0.0	-0.3	+0.4
1933	-0.3	-0.6	-0.3	+0.9	-0.6	-0.7	-0.3	-0.2	+0.5	-0.8
1934	-2.8	-0.8	-0.3	+0.6	+0.9	+1.2	+1.0	+0.4	+1.3	+0.4
1935	+0.6	+0.1	+0.4	+2.1	+1.7	0.0	+1.5	+0.7	-0.5	+0.3
1936	+1.4	+0.9	+0.7	-1.8	-0.6	-0.1	-0.7	-0.2	-0.8	-1.8
1937	-0.2	-0.7	-2.7	-2.6	-1.9	+0.1	+0.1	+0.4	0.0	+0.4
1938	+0.8	0.0	-0.4	-1.0	-1.6	+0.4	+0.3	-0.2	-0.2	+1.2

An attempt is made below to summarise the sequences which have occurred between weather types and ice conditions, with a note of such connections as appear to exist between them.

TYPE IA. The excess of IA in spring 1934 occurs with low ice in the Barentz Sea and relatively high ice in the Greenland-Iceland area. The greatest excess is in March, and in the next month, April, the ice is unusually low in the Barentz Sea, whilst it is uncommonly high in the westerly area. Similarly, an excess of IA in June - July 1930, July 1931, July 1933 and June 1938 is followed in all cases by a comparatively rapid fall in the Barentz area.

Therefore, broadly speaking, the wind circulation associated with type IA appears to have the effect of driving the ice away from the Barentz Sea north-eastwards and leading to a return flow along the Iceland-Greenland coastal region.

TYPE IB. An excess of IB appears to be associated with relatively high ice in the Barentz Sea. This was the case in spring 1932, and in 1936, whilst in the following years a strong fall in the ice was associated with a deficiency of IB. Brooks and Quennell[13] state that much ice in spring and summer in the Greenland-Iceland region is associated with high pressure

/over

over that area, and it can be noted that an excess of IB is accompanied by a rise in the ice in both areas.

TYPE ID. An excess of ID in 1931 and 1932 occurs with a rise in the yearly average ice, most marked in the Barentz Sea, whilst an excess of ID in summer 1935 is associated with fairly high ice, especially in the Barentz Sea.

According to Weise[11], heavy spring and summer ice in the Greenland Sea should lead to higher pressures in the Greenland-White Sea areas and lower pressures over the British Isles in autumn, i.e., to an excess of types IB and ID, and a deficiency of types II and III. There was heavy ice in this area in 1934, 1935, 1937 and 1938, and the autumn distribution of the above types was on the whole in accord with Weise's conclusions:-

	1934	1935	1937	1938	Totals
IB	+8	+8	-7	+1	+10
ID	+5	+6	+2	+15	+28
II	+6	-20	+5	-15	-24
III	-3	-2	0	-5	-10

Brennecke[12], investigating the variation of the ice off Iceland, found that in the spring of ice-poor years low pressure prevailed over Iceland and high pressure over Scandinavia. It is not possible to express these conditions exactly in terms of the pressure types, but on the whole they should imply higher frequencies of ID and III and possibly IB. The main ice-poor years in the Greenland-Iceland region were 1930, 1933 and 1936, and the distribution of the types in the springs of those years were as follows:-

	1930	1933	1936	Totals
IB	+3	+4	+2	+9
ID	-3	+9	0	+6
III	+1	+10	-3	+8

TYPE II. Generally speaking, type II is associated with high ice in the Greenland-Iceland area, whilst there is below the average in the Barentz Sea. An excess of type II in the winters of 1934 and 1938 is followed by comparatively high ice in the Greenland-Iceland area, whilst there is below the normal in the Barentz Sea. The deficiency of type II in early 1936 leads to high ice in the Barentz Sea with a sharp fall in July and August following an excess of II in May.

Brooks and Quennell[13] conclude that much ice in the Greenland-Iceland area is followed in winter by low pressure over the British Isles. The largest winter deficiency of type II is in 1938 - 39 following high ice in the region, whilst deficiencies in the winters of 1932-33, 1934-35 and 1935-36 also follow high ice in the Greenland-Iceland region. The comparatively large deficiency in winter 1936-37, however, is preceded by low ice, with every month below the average.

TYPE III. This type shows comparatively small anomalies and therefore the influence is not well marked, but an excess appears to be associated with low ice in the Barentz Sea.

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SUMMARY

The basic type is taken as the presence of the principal anticyclone in the Azores region and usually also across to Bermuda, the other types as variants produced by an extension or a displacement of this sub-tropical High Pressure belt in some direction, or by the presence of other high pressures.

Persistence of High Pressures.

The Azores High, with or without other high pressures of less importance (type I), is in evidence on nearly two thirds of the days of the year. In discussing the incidence and persistence of high pressure in other areas, special attention is devoted to types II and III, i.e., the main high pressure in the latitude of the British Isles, and north of latitude 60°N. respectively. It is found that type II, when persistent, has usually developed from the Azores High or arisen actually as a northward displacement of that High, and that type III is mostly a development of IB (the type where, in addition to high pressure over the Azores, there is high also in the latitude of Greenland), though in Winter there is a chance of type III developing from or into type II.

Atmospheric Circulation.

The possibility that displacements in latitude of the main high pressure system might show a tendency to recur at regular intervals is examined with results, rather inconclusive, which are shown in diagrams.

A daily zonal circulation index, taken as the difference between the highest pressure in the Atlantic south of latitude 50° N. and the lowest in the Icelandic area, is also examined. Apart from a strong seasonal variation, the series shows some tendency to recurrence of maxima and minima at intervals of about six days, or multiples thereof.

It is found that years which had the most frequent occurrence of very low pressures in the Icelandic area are for the most part those characterized also by very high pressures in the Southern N. Atlantic.

Seasonal and Monthly Sequence.

As regards particular years and seasons, high frequency of type II in Spring tends to be followed by infrequency in Autumn and vice versa, with a rather high correlation. Further, type II has frequency of occurrence above the normal in months both preceding and following abnormally warm months. That an excess of type II should follow a warm month is in accord with Brooks' conclusion that a high positive pressure anomaly in one month over the Azores will travel N.E., to appear as a high positive anomaly over the British Isles in the next month.

Weather of the Types.

An examination is made of the chief weather characteristics at Kew and Valentia with different weather types for cases when the same type lasted for three days or longer.

Movement of Depressions.

Although in general the tracks of depressions associated with the various types show great complexity, there are some fairly well defined tracks shown in some seasons. Type II, as might be expected from the frequency with which the anticyclone is over the British Isles, shows the chief tracks to the N. of Scotland and along the English Channel, especially in autumn and winter. Type IA shows congestion of tracks in certain fairly well defined areas. Type ID shows in all seasons an increased number of tracks from S.W. (as compared with IA) with fewer from S. and N.W. and, in spring and summer some from between S.E. and N.E. Type III shows concentrations of tracks from N. along the Scandinavian coast, from W. across S. England and from S.W. across France.

Arithmetic and vector means of speeds of travel of depressions and also frequencies are given in tables.

The arithmetic mean speeds of travel show remarkably little variation from season to season and from type to type, though individual cases differ widely. The depressions in ID and II travel fastest in all seasons except summer. The mean speeds by types and seasons range from 22 m.p.h. for IC in autumn and winter and IB in summer to 32 m.p.h. for II in winter. The mean of all is 26 m.p.h.

The seasonal vector means show an almost constant direction throughout the year for IA and ID (W. and W. by S. respectively) with high speeds, while II and III show an almost constant seasonal mean with more variable direction.

Relation to Ice Conditions.

A comparison of frequency of the various weather types with concurrent or subsequent ice conditions in the Barentz Sea and Greenland-Iceland region suggests that, broadly speaking, the effect of the basic type IA is to drive the ice from the Barentz Sea north eastwards and cause a return along the Iceland-Greenland Coastal region, whilst conversely excess of type IB brings the Barentz Sea ice limit southward.

As regards the relation of ice conditions to subsequent weather conditions, the data examined support on the whole Weise's conclusion that heavy spring and summer ice in the Greenland Sea leads in autumn to higher pressures in the Greenland-White Sea areas and lower pressures over the British Isles; in four out of five instances they agree with the conclusion of Brooks and Quennell that these ice conditions are followed also in winter by low pressure over the British Isles.

ACKNOWLEDGMENT.

The work described in this paper was carried out under the direction of Dr. A.H.R. Goldie, to whom I should like to express my thanks.

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TABLE 1 - CLASSIFICATION OF WEATHER TYPES.

JANUARY	1931	1932	1933	1934	1935	1936	1937	1938	1939	1940	1941
1	IB	ID	ID	ID	IB	IB	ID	II	III	IB	IA
2	IB	ID	ID	ID	ID	IB	ID	II	III	IB	ID
3	IB	IV	ID	ID	II	IB	ID	II	III	IA	II
4	IA	IV	ID	ID	II	IB	ID	II	III	IA	II
5	ID	ID	ID	IA	II	IB	ID	II	III	IA	II
6	II	ID	ID	ID	II	IB	ID	IB	IB	IA	II
7	II	IA	ID	ID	II	IB	IA	IA	ID	ID	II
8	II	IA	ID	ID	ID	IB	ID	IA	IV	II	II
9	II	IA	ID	ID	ID	IB	II	IA	ID	II	II
10	II	ID	II	ID	ID	IV	II	IA	IB	II	II
11	II	ID	ID	IV	ID	IV	IV	IA	III	II	II
12	II	ID	ID	IA	IC	II	IV	IA	III	II	II
13	II	ID	ID	IA	II	IB	ID	IA	IV	II	II
14	II	ID	ID	IA	II	II	ID	IA	IV	II	II
15	IA	ID	IA	IA	II	II	III	IA	IV	III	IC
16	IA	ID	IA	IA	II	IB	III	IA	ID	III	IC
17	IB	ID	IA	IA	II	II	III	IA	ID	III	IA
18	IA	IV	IA	IA	II	IB	III	IA	ID	III	IA
19	IB	ID	IA	IC	II	IB	III	IA	IA	III	IA
20	IB	ID	ID	ID	II	IV	III	IA	ID	III	IA
21	IB	ID	ID	II	II	IV	III	ID	ID	III	III
22	ID	ID	ID	II	II	IV	III	ID	IA	IB	IB
23	ID	II	ID	ID	II	IB	III	ID	ID	IB	IA
24	ID	II	II	ID	II	IB	IV	ID	IA	IB	IB
25	IA	II	II	ID	IA	IB	IV	IA	IC	III	IB
26	IC	II	II	ID	IC	IB	IV	IA	IB	III	IA
27	ID	II	II	IA	IA	IB	III	IA	III	III	IB
28	IA	II	II	II	ID	IB	III	IA	III	III	IB
29	IC	II	IA	II	ID	IB	III	IA	III	III	IA
30	IC	II	IA	II	IA	IB	III	IA	III	III	IB
31	ID	II	IA	II	IA	IB	III	IA	III	III	IB
FEBRUARY											
1	IA	II	ID	II	IA	IB	IV	IA	II	III	IC
2	IA	II	ID	II	IA	IB	IV	IA	II	III	IC
3	ID	II	IV	II	IA	IB	IV	IA	ID	III	IC
4	ID	II	ID	ID	IA	IB	IV	II	ID	III	ID
5	ID	II	ID	II	IA	II	IA	II	IV	III	IA
6	IA	IV	ID	II	IA	II	IA	ID	IV	III	IA
7	ID	IV	IB	II	II	II	IA	II	ID	III	IA
8	ID	II	IB	II	II	II	IA	V	ID	III	IA
9	ID	III	ID	II	II	III	IA	ID	IV	III	IA
10	ID	III	IA	IV	II	II	IA	IA	IV	III	IA
11	ID	II	II	II	ID	II	IA	II	IV	III	IA
12	IA	II	II	II	ID	II	IA	II	IA	III	IA
13	IC	II	II	II	IA	II	IB	II	IA	II	IA
14	IA	II	II	ID	IA	III	ID	II	ID	II	IA
15	IB	II	II	II	IA	III	ID	II	IA	III	IA
16	IB	II	II	II	IA	III	ID	II	IA	III	IB
17	IC	II	II	II	IA	III	IA	II	IA	III	IB
18	ID	II	III	II	ID	III	IA	II	IA	III	IB
19	ID	II	III	II	ID	III	ID	II	ID	IA	IB
20	ID	II	III	II	IA	III	IB	II	IA	IA	IB
21	ID	II	III	II	IA	IB	III	II	ID	IB	IB
22	ID	II	III	II	IA	IB	III	II	IA	IB	IA
23	ID	II	III	ID	IA	IC	III	II	IA	IB	IA
24	ID	II	III	IB	IA	IB	III	II	IB	IB	IA
25	ID	II	III	IC	IA	IB	III	IV	IB	IA	IB
26	ID	II	III	III	IA	IB	III	IV	ID	IA	IB
27	IB	II	III	IC	IA	IB	IB	ID	ID	IA	IB
28	IB	II	III	IA	IB	III	IB	ID	IA	IA	IB
29		II				III				ID	

TABLE 1 - CLASSIFICATION OF WEATHER TYPES.

MARCH	1930	1931	1932	1933	1934	1935	1936	1937	1938	1939	1940
1	II	IB	II	III	IA	IB	IC	IC	IA	IA	II
2	II	IB	IB	III	IA	IB	IA	IB	IA	IV	II
3	IV	III	II	III	IA	IB	IA	IV	II	IV	II
4	IV	III	II	III	ID	ID	IA	IV	II	ID	II
5	IV	III	II	III	IA	ID	IB	IV	II	ID	III
6	IV	III	II	III	IA	II	IB	III	II	ID	III
7	IV	III	III	IB	IA	II	IB	III	II	IA	III
8	IA	III	III	IA	IA	II	IC	III	IA	IA	III
9	IA	III	III	ID	IA	II	IB	III	IA	IA	II
10	IA	III	III	ID	IA	II	IA	III	II	II	IV
11	IA	III	II	ID	IA	III	IA	III	II	II	IV
12	IA	III	II	ID	IB	III	IA	III	II	II	III
13	IA	III	II	ID	IA	III	IA	III	II	II	III
14	IB	III	IV	IB	IA	III	IB	III	ID	II	III
15	III	III	II	IB	IA	IV	II	III	ID	II	III
16	III	II	II	IB	IA	IV	II	IA	ID	II	III
17	III	II	IB	IB	IA	IB	II	IA	ID	II	III
18	III	IV	IB	IB	IA	IB	II	IB	ID	II	III
19	IB	IV	ID	IB	IA	IB	II	IB	ID	II	ID
20	IB	IV	ID	IB	IA	IB	II	IB	ID	II	ID
21	III	IV	ID	IA	IA	ID	II	II	IV	IA	IB
22	IB	IV	ID	IB	IA	IB	III	II	IA	IB	IB
23	IB	IV	ID	II	IA	ID	III	IB	IA	IB	IB
24	ID	II	IB	ID	IA	IB	III	III	IA	IA	IA
25	II	II	III	ID	ID	II	III	III	IA	IB	IA
26	ID	II	IA	ID	IA	II	III	III	IA	III	IB
27	ID	II	ID	ID	ID	II	III	II	IA	III	IB
28	IB	III	ID	ID	IA	II	II	II	IA	III	IB
29	IA	II	ID	IA	IB	II	ID	II	IA	IB	IA
30	IA	III	IC	IA	IB	II	IV	II	IA	IB	IA
31	IA	III	IC	ID	IB	II	ID	II	IV	IB	IA
APRIL											
1	III	II	IB	IC	III	II	IV	II	II	IB	IA
2	III	IB	IB	II	III	II	III	III	II	IB	IA
3	III	IB	IC	II	III	II	II	IV	II	IB	IA
4	III	IC	IC	ID	III	II	II	III	II	IB	IA
5	III	ID	II	II	III	II	II	III	II	IB	IA
6	III	IA	IB	II	III	IB	II	III	II	IB	ID
7	III	IA	IB	II	III	IB	II	III	II	IA	ID
8	ID	II	IB	II	III	IB	II	III	II	ID	II
9	IA	ID	IB	II	III	IB	II	IB	II	IB	II
10	IC	ID	IB	IV	III	IB	II	IB	II	IB	II
11	IA	ID	IB	ID	III	IB	IB	IV	II	IB	II
12	IA	II	IC	IB	III	IB	IB	IB	II	IB	II
13	IA	II	ID	II	III	IB	III	IB	II	ID	II
14	IA	II	ID	II	III	IB	IB	IA	II	ID	II
15	IC	II	IB	II	III	IB	III	IA	II	ID	II
16	IC	II	II	II	IB	IC	III	IA	II	IA	IA
17	IC	IC	II	III	IB	IB	IB	IB	II	IA	III
18	IC	IC	IC	III	IB	IB	IB	IB	II	II	III
19	III	II	IB	III	IC	IB	IB	IB	II	II	III
20	III	II	IB	III	IA	IB	IB	IB	II	II	IA
21	II	II	IC	IB	IB	IB	IB	IA	II	II	IA
22	III	II	IB	ID	IC	III	IC	II	II	IC	IB
23	IB	ID	IB	ID	IA	II	IB	II	II	IA	IB
24	IB	ID	IB	IA	IA	II	IB	IB	II	IA	IA
25	IB	IB	IB	ID	IA	II	ID	IB	II	IA	IA
26	IB	ID	IA	IB	IC	II	IB	ID	II	IC	IA
27	IB	IB	IA	IB	IC	II	IB	ID	II	II	IA
28	IB	IB	IA	IC	ID	II	V	II	II	II	IA
29	IA	IB	IB	III	ID	II	ID	II	II	II	IA
30	IA	IC	IC	II	ID	II	V	II	II	II	IA

TABLE 1 - CLASSIFICATION OF WEATHER TYPES.

MAY	1930	1931	1932	1933	1934	1935	1936	1937	1938	1939	1940
1	IB	IC	III	IC	ID	II	II	II	II	IB	IC
2	IB	IC	III	III	IA	II	IB	II	III	IB	IA
3	IB	IC	III	III	IA	II	IB	IA	III	ID	IA
4	IB	IB	III	III	IB	IA	IB	IA	III	IB	IA
5	IB	IB	III	III	IB	IC	IB	ID	III	IB	ID
6	IB	IB	III	III	IB	V	IB	II	II	IB	II
7	IB	IB	III	IB	IB	II	IB	ID	II	IB	II
8	IB	IB	IB	IB	ID	II	IB	IB	III	III	II
9	IB	ID	IB	IB	ID	II	IB	IB	III	III	II
10	IA	ID	IB	IB	II	II	ID	IA	III	III	II
11	IA	ID	IB	IB	II	II	ID	IB	ID	III	II
12	IA	IA	ID	IB	II	II	ID	IB	ID	II	II
13	IA	IA	IB	IB	II	II	ID	IB	ID	II	II
14	IA	IA	IB	III	II	II	ID	IB	IA	II	IA
15	IA	IA	IB	III	II	II	ID	IB	IA	II	IB
16	ID	IA	IB	III	IB	II	III	IA	IB	II	IB
17	ID	IA	IB	III	IB	IC	II	IA	IB	II	IB
18	IA	IA	IB	III	IB	IC	II	IB	IB	IB	IB
19	II	IC	ID	III	IA	IB	II	IB	IB	IA	IB
20	II	IB	IA	III	IA	III	II	III	IB	IB	IB
21	II	IB	IB	III	ID	II	II	III	ID	IA	IB
22	II	IB	IB	IB	II	II	II	III	ID	IA	IB
23	ID	IB	IB	IB	II	II	II	III	ID	ID	IB
24	IC	IB	IC	II	II	II	II	III	IA	ID	IB
25	IB	IB	IC	IB	II	II	II	III	IB	II	IB
26	IA	IB	IC	IB	II	II	II	ID	IB	ID	II
27	IB	IB	IC	IB	II	II	II	II	IB	ID	IA
28	IB	IB	IB	IB	II	IB	II	II	IA	II	IA
29	IB	III	IB	IB	II	IB	II	II	IA	II	IA
30	IB	III	IB	IB	II	IB	II	ID	IA	II	IA
31	IB	III	IB	IB	II	IB	II	ID	IA	II	ID
JUNE											
1	IB	IB	IA	IB	II	IB	II	IA	IA	II	ID
2	IB	IB	IB	IB	II	IB	II	IC	IA	II	ID
3	IB	III	IB	IB	II	IB	IC	IA	IA	II	ID
4	ID	III	II	II	II	IB	IC	ID	ID	II	II
5	II	III	II	ID	II	IB	ID	IA	ID	II	ID
6	II	III	II	III	III	IB	IB	ID	ID	ID	ID
7	II	III	II	III	III	IB	IB	ID	ID	ID	ID
8	ID	III	ID	ID	III	IB	IB	ID	ID	II	ID
9	IA	III	ID	II	III	IB	IB	ID	ID	II	IA
10	IA	III	ID	ID	III	IB	IB	ID	ID	ID	IA
11	IA	III	ID	ID	IB	IB	IB	IA	ID	IB	IA
12	IA	III	IB	ID	II	IB	IB	II	II	IC	IA
13	ID	IB	ID	IB	IB	IB	IB	II	II	IA	IA
14	II	IB	ID	IB	II	IB	IB	II	II	IB	IA
15	II	IB	ID	IA	II	IB	IB	II	II	IB	IA
16	II	IB	ID	IB	II	IB	IB	II	II	IB	IA
17	ID	IB	II	IB	ID	ID	IB	II	II	IB	ID
18	ID	IB	IB	IB	ID	ID	ID	II	II	IB	ID
19	IA	IC	IB	IB	IA	ID	ID	II	IA	IB	ID
20	IB	ID	IB	IB	IA	ID	ID	II	IA	IB	ID
21	IB	ID	ID	IB	IA	ID	IB	II	IA	II	ID
22	IA	ID	II	IB	IA	II	ID	II	IA	III	IA
23	IA	IA	II	IB	IA	ID	IC	II	IA	III	IB
24	IA	ID	ID	IC	IB	ID	II	II	IA	III	IB
25	IA	ID	IB	IC	IB	ID	ID	II	IA	III	IB
26	IA	ID	IB	IC	IB	IB	ID	IA	IB	IB	IB
27	IA	II	ID	IC	IB	ID	II	IA	IA	ID	IB
28	IA	II	ID	IA	IB	ID	II	IA	IA	II	ID
29	IA	II	IA	IA	II	ID	II	IA	IA	ID	IA
30	IA	II	IB	IA	II	ID	IA	IA	IB	ID	ID

TABLE 1 - CLASSIFICATION OF WEATHER TYPES.

JULY	1930	1931	1932	1933	1934	1935	1936	1937	1938	1939	1940
1	IA	ID	IA	ID	II	ID	IA	IB	IB	IB	ID
2	IA	ID	IA	ID	II	ID	IA	II	IA	IB	IA
3	IA	IA	IA	II	II	ID	IA	II	IB	ID	IA
4	IA	IA	IB	II	II	ID	IA	IA	IA	ID	IA
5	IA	ID	IA	II	IB	II	ID	ID	IA	ID	IA
6	IA	IB	IA	III	III	II	IA	IB	IB	ID	IA
7	II	IB	ID	III	II	II	IA	IA	IC	IB	IA
8	II	IB	ID	IB	II	II	IA	IB	IB	IB	IA
9	II	IB	ID	ID	II	II	IB	IB	IB	IC	IA
10	II	IB	IA	IB	ID	ID	IB	IB	IB	II	IA
11	II	IA	IC	IA	IA	ID	IB	IB	IA	II	IB
12	II	IA	IC	IA	IB	ID	IB	IB	IA	ID	IB
13	IB	IA	IB	IA	IB	ID	IB	IB	IA	IB	IB
14	IB	IA	IB	IA	IB	ID	IB	IB	IA	IC	IC
15	IB	IA	IB	IA	IB	ID	IB	IA	IC	IC	IA
16	IB	IA	IC	IA	ID	IA	IB	IB	II	IA	IA
17	IB	IA	IC	IA	IB	IA	IB	ID	II	IB	IA
18	IB	IA	II	ID	IB	IA	IB	II	ID	IC	IA
19	IA	IA	IA	ID	IB	IA	IA	ID	IA	IA	IB
20	IA	IC	IA	ID	IC	IA	IA	ID	IB	IB	IB
21	II	IA	IA	IB	IB	IA	IA	ID	IB	IB	IB
22	II	ID	IC	IB	IA	IA	IA	ID	IA	IA	IA
23	II	ID	IA	IB	IA	ID	IB	ID	IA	IB	IA
24	IA	ID	IA	ID	IB	ID	IB	ID	IA	IB	IA
25	IA	IA	IB	IA	IA	ID	IA	II	IA	IB	IA
26	IB	IA	IA	IA	IA	ID	IB	II	IA	IB	IA
27	IA	IB	IA	IA	IB	IA	IB	II	IB	IB	IA
28	IA	IB	IA	IA	IA	IA	IA	IA	IB	IB	II
29	IA	IB	IA	IA	IA	II	IC	IA	ID	IA	II
30	IA	IA	IA	IA	IA	II	ID	IB	ID	IA	II
31	IA	IA	IA	IA	IA	II	IA	IB	ID	ID	II

AUGUST											
1	IA	IA	IA	IB	IB	II	IA	II	ID	ID	II
2	IA	ID	IA	II	IB	II	IA	III	ID	ID	II
3	IA	II	IA	II	IB	II	IB	III	ID	IA	ID
4	IB	II	ID	II	IB	II	IB	III	ID	ID	ID
5	IB	II	ID	II	IB	II	ID	III	ID	IB	ID
6	IB	II	ID	IA	IA	II	ID	IC	ID	IB	IA
7	IB	II	II	IA	IA	ID	II	ID	ID	IB	IA
8	IB	IC	II	IA	IA	ID	ID	ID	ID	IB	IA
9	IB	IC	II	IA	IA	IA	ID	ID	ID	IB	IA
10	IA	IC	II	IA	IA	IA	ID	IA	ID	IB	IA
11	IA	II	IA	IC	IA	IA	ID	IA	ID	ID	IA
12	IA	ID	IA	ID	IA	IA	IB	IA	IA	ID	IA
13	IA	IA	IA	II	IA	IA	IB	IB	IA	ID	IA
14	IA	IB	IC	IA	IA	IA	IB	IB	IA	ID	IA
15	IA	IB	ID	IA	IA	IA	IB	IB	IA	ID	IA
16	IB	IB	IA	IA	IA	ID	ID	IB	IA	II	ID
17	IB	IB	ID	IA	IA	ID	ID	IB	IA	II	II
18	IB	IB	ID	IA	IA	IA	ID	ID	IA	ID	II
19	IB	IB	ID	IA	IA	ID	ID	ID	IB	ID	ID
20	IB	IB	II	IA	IA	ID	IB	ID	IB	ID	ID
21	IB	IA	II	IA	IA	IB	ID	ID	IA	IB	IA
22	IB	IA	II	IA	IA	IB	IB	ID	IA	IB	IA
23	IB	IA	II	IA	IA	IB	II	ID	IA	IA	IC
24	IB	IA	II	IB	IA	IB	II	ID	ID	ID	ID
25	II	IC	II	II	II	ID	II	ID	ID	IB	ID
26	II	II	II	II	II	IA	II	II	ID	III	II
27	II	II	ID	II	II	IB	II	II	IB	III	II
28	II	II	ID	II	III	IB	II	III	IB	IB	II
29	II	II	IA	ID	III	IB	II	III	IC	IB	II
30	II	II	IA	ID	III	IB	II	ID	IB	IB	ID
31	II	ID	IA	ID	IB	IB	II	II	IB	IA	ID

TABLE 1 - CLASSIFICATION OF WEATHER TYPES.

SEPTEMBER	1930	1931	1932	1933	1934	1935	1936	1937	1938	1939	1940
1	II	IB	ID	IA	IB	IB	IB	ID	IB	III	II
2	II	IB	ID	ID	IB	IB	IB	ID	IB	II	II
3	II	IB	ID	II	IB	IA	IB	ID	ID	IA	ID
4	II	IC	IA	II	ID	IB	IB	ID	ID	IA	IA
5	IB	IC	ID	III	II	IB	IB	ID	ID	ID	IA
6	IB	II	IA	ID	III	IA	IB	ID	ID	II	ID
7	IB	II	IA	II	III	III	IB	ID	IB	II	IA
8	IB	IC	IA	II	IA	II	IB	II	ID	ID	IA
9	IB	IC	IA	II	IA	II	IA	IC	ID	ID	IA
10	IB	IC	IA	II	ID	II	II	IC	ID	IA	II
11	IB	ID	IA	II	ID	ID	II	ID	IB	IA	II
12	III	IC	IB	IB	ID	ID	II	ID	II	IA	II
13	IA	ID	IA	IB	II	ID	II	IA	II	IA	IA
14	II	II	II	II	II	ID	ID	IA	ID	IA	IA
15	II	II	II	II	II	ID	II	ID	II	II	IA
16	IB	II	ID	II	ID	IA	III	ID	II	II	IA
17	IB	II	ID	ID	ID	IA	II	IA	ID	II	IA
18	IA	II	IB	IA	ID	IA	II	IA	ID	II	IA
19	IA	II	II	IA	ID	ID	II	IA	ID	II	IA
20	IA	II	II	IA	ID	ID	II	IC	IB	II	IA
21	IB	II	II	IA	IB	IA	IV	IA	IB	II	IA
22	ID	II	II	IA	IB	ID	II	IA	IV	II	IA
23	ID	II	II	IA	IB	ID	II	ID	IV	III	IA
24	IA	II	II	IC	IB	ID	III	IA	IV	II	II
25	II	II	IB	IA	ID	IB	III	ID	ID	II	II
26	II	II	IC	ID	ID	IB	II	IV	ID	III	II
27	II	II	IC	III	ID	IB	II	IV	ID	III	II
28	II	ID	ID	III	IV	IB	II	IV	IV	II	II
29	II	ID	IA	II	ID	ID	II	II	IV	IA	II
30	II	ID	III	II	IB	ID	II	II	ID	IB	II
OCTOBER											
1	II	IV	II	II	ID	IA	II	ID	ID	III	II
2	II	II	IC	II	IB	IA	II	ID	ID	IA	II
3	II	II	IC	II	IA	IA	II	ID	ID	II	ID
4	ID	II	ID	II	IA	IA	IB	II	ID	II	ID
5	IA	ID	IA	II	IB	IC	III	II	IB	IB	ID
6	IA	IV	IA	II	ID	IA	III	II	IB	IA	IA
7	IB	ID	IA	IV	ID	IA	II	III	ID	III	IA
8	IA	ID	IA	IV	ID	IA	II	III	ID	III	IA
9	IB	IV	IA	IV	II	ID	II	II	IB	III	IA
10	ID	ID	IA	IV	II	ID	II	II	IB	III	IA
11	IV	II	IA	IV	II	ID	II	II	IB	III	IA
12	IV	II	IA	ID	II	ID	II	II	ID	III	ID
13	IV	II	IA	IA	II	ID	IA	II	ID	IB	IV
14	IV	II	IA	ID	II	ID	IA	IA	ID	IB	IV
15	IV	II	IA	ID	II	ID	IA	II	ID	IA	III
16	IV	II	IA	ID	II	ID	IA	II	IB	IA	IV
17	ID	II	IA	ID	II	ID	IA	ID	IA	IB	IA
18	ID	II	IA	IV	IA	ID	IA	IV	ID	ID	IA
19	IV	II	IA	IV	IA	IC	IA	ID	ID	ID	IB
20	IV	III	IA	III	ID	IC	IA	ID	II	II	IB
21	IA	II	IA	III	IV	IA	IA	ID	II	ID	IB
22	IA	III	IA	III	IV	IA	ID	IV	ID	II	III
23	IA	III	ID	III	IV	IA	ID	IV	ID	IA	III
24	IA	IC	IA	III	IV	IB	ID	IV	ID	IC	III
25	II	II	IA	III	IV	IB	ID	IV	ID	IC	IB
26	IA	II	IA	II	ID	IB	IA	IV	II	IC	IB
27	IA	II	IA	II	ID	IB	ID	IV	II	II	IB
28	IA	IC	IA	II	IB	IB	IA	IV	IA	IB	ID
29	IA	IB	IA	II	IB	IB	ID	IV	IA	IC	IA
30	ID	ID	IB	II	IB	IB	ID	IV	IA	IB	IA
31	ID	II	IA	II	II	III	ID	IV	ID	III	IA

TABLE 1 - CLASSIFICATION OF WEATHER TYPES.

NOVEMBER	1930	1931	1932	1933	1934	1935	1936	1937	1938	1939	1940
1	IA	IV	IA	II	II	IV	ID	IV	IA	IB	IA
2	IA	IV	II	II	IB	IV	IA	IV	IA	III	IA
3	IA	ID	IV	II	IB	IV	ID	ID	IA	III	IA
4	IA	ID	IC	II	IB	IV	ID	ID	IA	IV	IA
5	IA	ID	II	II	II	IV	ID	ID	IA	IV	IB
6	ID	ID	II	II	IC	IV	IV	IV	IV	IV	IB
7	ID	ID	IA	II	II	IV	IV	IV	IV	ID	IB
8	ID	ID	IA	IA	IA	IB	IV	II	IV	IV	ID
9	ID	ID	III	IA	IC	IV	IA	II	IV	ID	IA
10	II	ID	IV	IA	IC	IV	ID	II	IV	ID	IA
11	II	ID	IV	IA	IA	IV	ID	II	IV	IB	IA
12	II	ID	III	IA	IA	IV	ID	II	IV	IB	IA
13	IV	ID	III	IA	IA	IV	IV	II	IV	IB	IA
14	IB	ID	II	IB	IB	ID	ID	II	ID	ID	IA
15	IB	III	II	IB	IC	ID	ID	II	ID	ID	IA
16	II	III	II	IC	ID	ID	ID	IV	ID	IA	IA
17	II	III	II	IC	ID	IV	IA	IA	ID	IA	IA
18	II	III	II	III	ID	III	II	III	ID	IA	IA
19	ID	III	IV	II	II	IV	II	IB	ID	IA	IA
20	ID	ID	IV	II	II	IV	II	III	IA	IA	IA
21	ID	ID	IA	II	II	IV	II	IC	IA	ID	IA
22	ID	IV	IA	ID	II	IB	II	II	IA	ID	IA
23	IA	IV	IA	IA	II	IB	II	II	IA	ID	IA
24	IA	ID	IA	III	II	ID	IC	II	IA	ID	II
25	IA	III	IA	II	II	ID	ID	II	ID	ID	II
26	IA	ID	IA	II	II	ID	ID	II	ID	ID	II
27	IA	ID	IA	II	II	ID	II	II	ID	IA	II
28	IC	III	IA	III	II	ID	II	II	ID	IA	II
29	ID	ID	ID	II	II	ID	ID	ID	IA	IA	II
30	II	ID	ID	II	II	ID	IA	ID	IA	IA	II

DECEMBER	1930	1931	1932	1933	1934	1935	1936	1937	1938	1939	1940
1	II	ID	IA	III	IV	ID	IA	IA	IA	IA	II
2	II	ID	IA	III	ID	IB	IA	II	IA	IA	II
3	II	ID	IA	II	IA	IB	IA	III	ID	IA	II
4	II	ID	IA	II	IA	IB	IA	IB	ID	IA	IA
5	II	IA	IA	II	IA	IB	II	IB	ID	IA	IA
6	II	ID	IB	III	IA	IB	II	IB	IA	IA	IA
7	IA	ID	III	III	IA	IV	II	III	ID	IA	IA
8	IA	ID	II	II	IV	ID	II	II	IB	IA	IA
9	ID	II	III	II	ID	ID	ID	II	IB	IA	IA
10	ID	II	II	II	ID	II	ID	IB	IB	ID	IA
11	IB	II	II	II	IV	II	ID	IC	IB	IB	IA
12	IB	II	II	II	IV	II	IA	IA	IV	ID	IA
13	ID	II	IV	II	IV	II	IA	IA	III	V	ID
14	ID	II	ID	II	IV	ID	ID	IA	III	ID	IA
15	III	II	ID	IA	IV	ID	ID	IB	III	ID	IA
16	III	II	ID	ID	IB	IV	ID	IC	III	ID	ID
17	III	II	ID	II	IA	IV	ID	ID	III	II	ID
18	ID	II	ID	II	IA	IB	ID	ID	ID	II	IV
19	ID	II	IV	II	IA	IB	ID	IB	IB	III	IV
20	ID	II	ID	II	IA	IB	ID	III	IB	III	II
21	ID	II	ID	II	IA	IB	IV	II	III	II	II
22	II	II	ID	II	III	IB	ID	ID	III	II	III
23	II	II	ID	II	III	III	II	ID	ID	II	III
24	IA	II	ID	II	III	IV	II	ID	ID	II	III
25	IB	II	II	IA	III	III	II	II	ID	II	II
26	ID	II	ID	IB	III	III	II	III	ID	II	II
27	IB	II	ID	IB	III	III	IV	V	ID	II	II
28	IA	II	ID	IB	III	IV	IV	II	IA	III	II
29	ID	III	ID	IB	III	IV	ID	II	IA	III	IV
30	ID	IC	ID	IB	ID	ID	ID	II	ID	II	IB
31	ID	ID	IV	IB	ID	ID	ID	II	IB	IV	IA

TABLE 2 - AVERAGE NUMBER OF DAYS PER MONTH ON WHICH EACH TYPE HAS OCCURRED.

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
							A			1930-1941			
IA	6.1	6.4	6.4	3.8	4.0	5.4	12.4	8.9	5.7	7.5	7.0	5.3	78.8
IB	4.2	3.5	5.1	7.7	10.7	8.0	8.4	6.4	4.3	3.2	1.7	3.4	66.6
IC	0.8	0.7	0.5	2.3	1.3	0.9	1.4	0.8	1.1	1.0	0.9	0.3	12.0
ID	7.5	4.5	3.9	2.5	3.3	7.6	4.8	7.5	7.0	6.8	7.6	7.8	70.8
II	7.0	7.6	6.6	9.1	7.9	6.2	3.8	6.5	9.5	6.7	6.7	8.3	85.9
III	3.7	4.2	6.3	4.0	3.7	1.9	0.3	1.0	1.4	2.4	1.5	3.6	34.0
IV	1.6	1.4	2.2	0.4	-	-	-	-	0.9	3.4	4.3	2.2	16.4
V	-	0.1	-	0.2	0.1	-	-	-	-	-	-	0.2	0.6
							B			1896-1910			
IA	7.9	9.1	8.9	7.5	6.8	9.8	14.4	14.0	8.3	8.5	5.5	9.9	110.6
IB	1.7	1.2	4.0	5.9	4.4	4.1	2.1	3.0	2.9	2.6	2.9	2.4	37.2
IC	1.4	0.9	1.5	2.1	2.7	2.2	3.7	2.5	1.1	1.1	1.1	0.8	21.1
ID	10.0	5.0	5.3	3.3	4.2	4.8	5.8	4.1	5.9	5.4	8.7	8.9	71.4
II	6.8	7.3	6.3	4.7	8.4	5.2	3.9	4.9	8.5	8.7	8.0	5.8	78.2
III	0.1	1.5	2.9	3.1	2.6	2.1	0.2	0.8	0.8	1.8	1.5	0.7	18.1
IV	0.5	0.4	0.3	0.1	0.1	-	-	-	-	0.5	0.3	0.2	2.4
V	2.3	1.5	0.1	0.7	0.9	0.7	0.3	0.3	1.7	0.8	0.8	1.0	11.1
							C			1896-1910; 1930-1941			
IA	7.2	8.0	7.8	6.0	5.6	8.0	13.5	11.8	7.2	8.1	6.0	8.0	97.2
IB	2.7	2.2	4.5	6.7	7.1	5.8	4.7	4.4	3.5	2.8	2.4	2.8	49.6
IC	1.2	0.8	1.0	2.2	2.1	1.7	2.7	1.8	1.1	1.1	1.0	0.6	17.3
ID	8.9	4.8	4.7	3.0	3.8	6.0	5.4	5.5	6.4	6.0	8.2	8.5	71.2
II	6.9	7.5	6.5	6.5	8.0	5.6	3.9	5.6	8.9	7.8	7.5	6.8	81.5
III	1.7	2.6	4.3	3.5	3.1	2.0	0.2	0.9	1.0	2.0	1.5	1.9	24.7
IV	1.0	0.8	1.1	0.2	0.1	-	-	-	0.4	1.7	2.0	1.0	8.3
V	1.3	0.9	-	0.5	0.6	0.4	0.2	0.2	1.0	0.5	0.5	0.7	6.8
Unclassified	0.1	0.4	1.1	1.4	0.6	0.5	0.4	0.8	0.5	1.0	0.9	0.7	8.4

TABLE 3 - FREQUENCY OF OCCURRENCE 1905-18. APPROXIMATE GROUPING OF GOLD'S TYPES 1905-18, TOGETHER WITH THE DERIVED FREQUENCIES IN TERMS OF GOLDIE'S TYPES.

Gold	Goldie	Jan.	Feb.	Mar.	Apr.	May	Jne.	Jly.	Aug.	Sept.	Oct.	Nov.	Dec.	Yr.	Annual Mean
I, IA, II) III, IV) IVA, XV)	IA, IB	191	148	158	129	113	156	222	201	139	120	156	172	1905	136
IX, IXA) X)	IC	22	19	40	57	40	60	60	35	42	21	27	19	442	32
V, VA) X, XIV)	ID	58	91	78	65	43	52	40	67	54	70	77	93	788	56
VI, VIA,) VIIA, VIIB) VIIIB, IXB) XIA, XIIIA)	II	118	92	82	100	145	104	77	83	140	120	95	85	1243	89
VIIC,) VIII, VIIIA)	III	22	17	32	40	49	20	16	8	28	35	28	16	311	22
VII	IV	15	14	26	13	24	20	11	20	11	36	21	34	245	17

TABLE 4 - FREQUENCY OF RUNS OF GIVEN DURATION. March 1930 - February 1941.

		Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
IA	1- 5	23	18	17	16	17	19	37	18	24	22	18	18
	6-10	2	3	3	1	2	3	8	5	2	3	2	2
	11-15	1	1	1	-	-	-	-	-	1	-	1	-
	16 and over	-	-	-	-	-	-	-	1	-	1	-	-
	Total No.runs:	26	22	21	17	19	22	45	24	27	26	21	20
	Total No.days:	67	70	70	42	44	59	136	98	63	83	77	58
	Mean duration:	2.6	3.2	3.3	2.5	2.3	2.7	3.0	4.1	2.3	3.2	3.7	2.9
IB	1- 5	17	14	27	25	20	19	34	17	17	18	10	16
	6-10	2	1	1	4	6	3	4	4	2	1	-	1
	11-15	-	-	-	-	1	1	-	-	-	-	-	-
	16 and over	-	-	-	-	-	1	-	-	-	-	-	-
	Total No.runs:	19	15	28	29	27	24	38	21	19	19	10	17
	Total No.days:	46	38	56	85	118	88	92	70	47	35	19	37
	Mean duration:	2.4	2.5	2.0	2.9	4.4	3.7	2.4	3.3	2.5	1.8	1.9	2.2
IC	1- 5	7	6	4	19	8	6	12	7	7	7	8	3
	Total No.days:	9	8	5	25	14	10	15	9	12	11	10	3
	Mean duration:	1.3	1.3	1.3	1.3	1.8	1.7	1.3	1.3	1.7	1.6	1.3	1.0
ID	1- 5	24	23	17	17	19	32	22	30	36	33	24	34
	6-10	3	1	1	-	1	2	2	1	1	1	3	1
	11-15	-	-	-	-	-	-	-	1	-	-	1	-
	Total No.runs:	27	24	18	17	20	34	24	32	37	34	28	35
	Total No.days:	83	49	43	28	36	84	53	82	77	75	84	86
	Mean duration:	3.1	2.0	2.4	1.6	1.8	2.5	2.2	2.6	2.1	2.2	3.0	2.5
II	1- 5	10	12	18	17	12	21	13	16	26	19	14	18
	6-10	3	2	2	3	6	1	1	4	4	5	5	4
	11-15	2	1	1	-	1	1	-	-	1	-	1	-
	16 and over	-	1	-	1	-	-	-	-	-	-	-	1
	Total No.runs:	15	16	21	21	19	23	14	20	31	24	20	23
	Total No.days:	77	84	73	100	87	68	42	71	104	74	74	91
	Mean duration	5.1	5.3	3.5	4.8	4.6	3.0	3.0	3.6	3.4	3.1	3.7	4.0
III	1- 5	5	5	10	11	7	3	2	4	11	9	12	17
	6-10	3	2	4	1	3	1	-	-	-	2	-	1
	11-15	-	2	1	1	-	-	-	-	-	-	-	-
	Total No.runs:	8	9	15	13	10	4	2	4	11	11	12	18
	Total No.days:	41	46	69	44	41	21	3	11	15	26	18	40
	Mean duration:	5.1	5.1	4.6	3.4	4.1	5.3	1.5	2.8	1.4	2.4	1.5	2.2
IV	1- 5	8	7	9	4	-	-	-	-	5	10	16	16
	6-10	-	-	1	-	-	-	-	-	-	2	2	-
	Total No.runs:	8	7	10	4	-	-	-	-	5	12	18	16
	Total No.days:	18	15	24	4	-	-	-	-	10	37	47	24
	Mean duration:	2.3	2.1	2.4	1.0	-	--	-	-	2.0	3.1	2.6	1.5
V	1- 5	-	1	-	2	1	-	-	-	-	-	-	2
	Total No.days:	-	1	-	2	1	-	-	-	-	-	-	2
	Mean duration:	-	1	-	1	1	-	-	-	-	-	-	1

TABLE 5 - SUCCESSION OF TYPES. March 1930 - February 1941.

SPRING

Following Type	Preceding Types							
	IA	IB	IC	ID	II	III	IV	V
IA	64	8	17	16	3	2	7	
IB	15	70	26	14	5	8	7	
IC	6	5	33	3	1	1		
ID	9	6	4	50	4	3	11	33
II	3	3	7	12	80	10	11	66
III	1	6	11	1	5	75	14	
IV	1	1		3	2	1	50	
V			2	1				

SUMMER

Following Type	Preceding Types							
	IA	IB	IC	ID	II	III	IV	V
IA	57	13	23	16	6			
IB	22	68	11	11	2	19		
IC	5	4	26		1	3		
ID	13	9	20	61	16	6		
II	3	4	20	12	72	3		
III		1			3	69		
IV								
V								

AUTUMN

Following Type	Preceding Types							
	IA	IB	IC	ID	II	III	IV	V
IA	68	10	15	15	5	5	5	
IB	6	53	6	6	4	7	3	
IC	3	5	30	1	2	3	1	
ID	13	16	24	61	9	7	18	
II	6	6	21	9	73	27	4	
III	3	8	3	2	4	42	4	
IV		2		6	3	8	64	
V								

WINTER

Following Type	Preceding Types							
	IA	IB	IC	ID	II	III	IV	V
IA	67	11	25	15	3	1	5	
IB	8	60	10	7	2	5	9	
IC	3	5	20			2		
ID	16	8	35	61	4	3	29	66
II	4	3	5	9	80	12	9	33
III	1	9	5	1	6	73	5	
IV		4		6	4	4	43	
V						1		

TABLE 9 - VALUES OF PRESSURE TAKEN FROM M.O.M.394 (Western North Atlantic) and M.O.M.447 (Home Waters) PRESSURE READINGS 1921-1938.

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Seasonal Range	Yr. Mean
1. Maxm. pressure in Azores High	1025	1023	1021	1021	1025	1023	1025	1024	1023	1021	1023	1023	4	
2. Lat. of Azores High	35	35	35	30	30.5	29	35	37.5	35.5	32.5	34	36	8.5°	
3. Min. pressure Icelandic Low	995	995	1005	1007	1011	1009	1007	1009	1006	1003	999	999	16	
4. Lat. of centre of Icelandic Low in long. 25°-30° W.	63	62	63	61.5	58	61	65.5	62	62	62.5	60.5	63	7.5°	
5. Mean W-E Gradient	1.07	1.04	0.57	0.44	0.51	0.44	0.59	0.61	0.64	0.60	0.91	0.89	0.63	0.69

TABLE 12 - SEASONAL FREQUENCY OF TYPES. 1930-1941.

	IA	IB	IC	ID	II	III	IV	V
Winter	17.1	10.7	1.8	12.8	22.8	11.4	12.8	0.9
Spring	13.5	23.2	4.0	8.9	23.5	13.8	4.6	0.3
Summer	26.7	22.7	3.0	19.8	16.4	3.2	0.2	0.1
Autumn	19.3	8.8	3.0	17.7	23.2	5.4	13.2	0.2

TABLE 13 - SEASONAL FREQUENCY - DIFFERENCE FROM NORMAL

	TYPE I				TYPE II				TYPE III				TYPE IV		
	W.	Sp.	S.	A.	W.	Sp.	S.	A.	W.	Sp.	S.	A.	W.	Sp.	A.
1930		+14	-3	-1		-15	+6	+3		+1	-3	-4		0	+3
1	+23	-1	-7	-10	-6	-6	-1	+10	-8	+3	+7	+5	-9	+3	-4
2	-22	-6	0	+14	+31	-10	+3	-7	-8	-2	-3	-1	0	-4	-5
3	-2	+4	0	-23	-5	-10	-3	+16	+1	+10	+1	+7	+4	-4	0
4	-10	+11	-11	+6	+17	-7	+5	+6	-6	+1	+6	-3	-5	-5	-7
5	+7	-14	+4	+11	-2	+22	-1	-20	-3	-8	-3	-2	-2	-1	+12
6	+1	-9	+3	-5	-7	+9	0	+10	+3	-3	-3	-1	+4	+1	-4
7	-4	-12	-11	-13	-13	-4	+8	+5	+9	+11	+3	0	+7	+4	+9
8	-1	-18	+9	+6	+9	+20	-7	-15	-7	-8	-3	-5	-3	+6	+14
9	-9	+2	0	+1	-21	+7	-3	-4	+8	-7	+3	+9	+22	-2	-4
1940	-9	+3	+6	+14	-4	-1	-3	-2	+23	0	-3	0	-11	-2	-11
1	+18				-2				-7				-10		

	TYPE IA				TYPE IB				TYPE ID			
	W.	Sp.	S.	A.	W.	Sp.	S.	A.	W.	Sp.	S.	A.
1930		+11	+11	+5		+3	+4	+2		-3	-15	-6
1	-2	-5	-4	-19	+4	-1	0	-5	+18	+1	-5	+9
2	-13	-9	+2	+26	-11	+14	-9	-5	+3	+3	+4	-9
3	+2	-8	+6	-4	-7	+4	-3	-5	+5	+9	-5	-14
4	-6	+18	+6	-9	-4	-8	+1	+8	-1	+1	-16	+5
5	+16	-14	-9	-3	-8	+6	+3	+8	-1	-5	+13	+6
6	-17	-10	-10	-5	+31	+2	+12	0	-12	0	0	+2
7	-3	-8	-10	-9	-7	+1	-6	-7	+8	-2	+6	+2
8	+11	+5	+9	-7	-4	-16	-6	+1	-8	-3	+6	+15
9	-6	0	-18	-1	-3	+1	+12	+2	+1	+3	+4	+1
1940	+1	+15	+13	+29	-1	-5	-12	-1	-7	-5	+6	-11
1	+18				+7				-9			

TABLE 16 (a)-(g) - WEATHER ASSOCIATED WITH PRESSURE TYPES (0 signifies no observations, - no occurrences of the type. Figures in brackets are means after excluding thunder rain).

(a) Daily Rainfall

KEW

	J.	F.	M.	A.	M.	J.	Jy.	A.	S.	O.	N.	D.
IA	2.3	2.4	1.2	(1.8) 2.2	1.7	1.0	(1.5) 1.8	0.6	(2.1) 2.6	(2.3) 2.4	(4.0) 4.1	2.2
IB	2.7	2.0	2.2	2.3	(1.8) 2.2	(1.5) 2.3	(1.7) 2.4	(2.9) 4.4	(1.9) 3.0	0.6	1.1	1.4
ID	1.7	1.1	0.7	0.15	0.02	1.2	1.6	1.4	1.5	2.0	3.2	1.0
II	0.2	0.1	0.2	0.7	0.7	(0.7) 1.1	(0.2) 0.3	(0.1) 1.1	(0.9) 1.0	0.4	0.4	0.3
III	3.3	2.7	2.9	1.4	1.7	(1.1) 1.5	-	(3.0) 3.2	-	2.4	2.3	3.4
IV	3.6	1.7	0.7	-	-	-	-	-	1.7	(2.4) 2.7	4.0	4.7

TABLE 16 continued.

(a) Daily Rainfall.

VALENTIA

	J.	F.	M.	A.	M.	J.	Jy.	A.	S.	O.	N.	D.
IA	6.7	(4.5) 4.4	4.9	3.6	3.4	(3.3) 3.5	3.7	3.3	(8.2) 8.6	5.7	5.1	(5.4) 5.8
IB	(8.2) 8.0	3.8	(2.5) 3.1	4.6	3.4	4.5	(5.5) 5.4	6.1	4.6	2.5	6.4	(4.5) 4.5
ID	(5.9) 6.9	(4.8) 4.9	4.4	0.5	6.7	4.5	4.0	2.8	5.4	4.1	7.5	(3.7) 4.0
II	0.5	0.8	1.2	0.3	0.6	0.8	0.4	(3.3) 3.9	1.5	1.4	1.4	1.9
III	(14.0) 12.7	(4.7) 4.6	3.0	3.3	(2.8) 3.0	3.4	-	2.8	-	0.6	13.7	(12.4) 12.2
IV	(8.1) 9.8	8.1	1.9	-	-	-	-	-	8.5	9.4	(8.1) 8.3	9.0

(b) % day with Rain.

KEW

	J.	F.	M.	A.	M.	J.	Jy.	A.	S.	O.	N.	D.
IA	80	66	51	68	58	34	53	41	75	62	71	64
IB	78	68	48	68	54	57	56	51	67	42	56	65
ID	60	69	29	33	6	28	35	20	44	54	73	52
II	35	14	18	29	26	25	13	16	34	26	35	38
III	68	65	58	44	38	42	-	43	-	53	60	84
IV	78	71	21	-	-	-	-	-	33	62	71	100

VALENTIA

	J.	F.	M.	A.	M.	J.	Jy.	A.	S.	O.	N.	D.
IA	98	(86) 87	85	100	76	(72) 73	94	86	(89) 95	96	95	86
IB	96	100	(85) 86	89	67	86	87	93	88	77	100	91
ID	100	91	80	33	92	71	81	71	(82) 83	87	86	91
II	139	30	24	30	(32) 33	36	33	(61) 62	(40) 42	45	45	57
III	100	(91) 92	68	58	68	60	-	57	-	67	100	100
IV	100	100	71	-	-	-	-	-	80	85	(96) 97	100

(c) % frequency Thunder.

KEW

	J.	F.	M.	A.	M.	J.	Jy.	A.	S.	O.	N.	D.
IA	2	0	2	8	4	2	9	7	5	3	2	0
IB	0	0	3	4	9	18	5	15	18	0	0	0
ID	0	0	0	0	0	10	8	4	2	24	0	0
II	0	0	0	0	4	7	3	9	4	0	0	0
III	0	0	0	3	5	5	-	29	-	0	0	0
IV	0	0	0	-	-	-	-	-	0	4	0	0

VALENTIA

J.	F.	M.	A.	M.	J.	Jy.	A.	S.	O.	N.	D.
0	7	0	0	0	3	0	0	10	0	0	5
4	0	9	0	0	0	2	0	0	0	0	5
15	9	0	0	0	0	0	0	3	0	0	2
0	0	0	0	2	0	0	2	2	0	0	0
14	8	0	0	3	0	-	0	-	0	0	9
17	0	0	-	-	-	-	-	0	0	10	20

(d) % Possible Sunshine.

KEW

	J.	F.	M.	A.	M.	J.	Jy.	A.	S.	O.	N.	D.
IA	16	25	28	26	30	41	32	40	34	31	25	27
IB	11	20	31	30	36	37	36	40	32	24	24	15
ID	16	15	45	33	47	48	49	50	36	26	18	18
II	13	28	33	36	45	48	52	54	39	28	21	6
III	19	18	28	24	34	26	-	46	-	18	1	43
IV	23	23	22	-	-	-	-	-	48	31	20	10

VALENTIA

IA	14	20	27	31	43	27	21	28	33	18	22	13
IB	19	25	30	29	36	29	21	30	28	28	19	25
ID	12	18	26	43	29	29	27	27	27	24	22	16
II	27	35	36	50	51	47	43	31	45	28	26	19
III	10	25	32	33	40	31	-	43	-	41	15	19
IV	16	33	36	-	-	-	-	-	32	24	20	10

(e) % frequency Sunless days.

KEW

IA	37	25	13	12	13	4	6	2	10	14	24	21
IB	41	32	19	18	6	1	5	7	6	16	33	35
ID	51	22	11	17	0	7	0	5	9	17	36	50
II	49	30	10	9	5	4	3	0	10	14	37	76
III	42	40	17	24	21	16	-	0	-	47	60	63
IV	33	29	29	-	-	-	-	-	0	8	39	40

VALENTIA

IA	32	23	17	0	12	18	13	13	5	37	18	36
IB	37	9	14	9	7	16	13	17	18	15	66	30
ID	44	39	40	0	17	18	22	12	17	32	25	45
II	33	32	18	5	9	8	11	18	14	20	38	52
III	43	8	24	19	6	7	-	0	-	0	40	6
IV	17	25	14	-	-	-	-	-	17	23	18	20

(f) % frequency Fog.

KEW

IA

	J.	F.	M.	A.	M.	J.	Jy.	A.	S.	O.	N.	D.
IA	31	23	25	4	0	6	0	6	5	17	29	21
IB	30	12	19	7	8	4	0	7	12	21	33	35
ID	37	17	32	17	0	0	3	9	4	32	31	29
II	34	37	30	5	2	2	0	23	17	28	50	53
III	37	40	18	26	0	0	-	14	-	47	60	37
IV	33	0	29	-	-	-	-	-	17	19	23	0

VALENTIA

IA	5	0	3	0	0	3	0	0	0	9	0	0
IB	0	0	0	0	2	5	0	2	0	0	0	0
ID	0	9	6	0	17	8	3	18	0	0	0	0
II	0	0	3	0	2	2	11	4	12	0	0	0
III	0	0	2	13	0	0	-	0	-	0	0	0
IV	0	0	0	-	-	-	-	-	0	0	0	0

(g) % frequency Snow.

KEW

IA	0	4	4	0	0	0	0	0	0	0	0	0
IB	7	4	3	2	0	0	0	0	0	0	0	4
ID	2	9	0	0	0	0	0	0	0	0	0	9
II	4	14	7	2	1	0	0	0	0	0	0	0
III	11	17	18	3	0	0	-	0	-	0	0	0
IV	0	0	0	-	-	-	-	-	0	0	0	0

VALENTIA

IA	5	0	0	0	0	0	0	0	0	0	0	0
IB	0	18	5	4	0	0	0	0	0	10	0	0
ID	0	0	0	0	0	0	0	0	0	0	0	0
II	0	2	0	0	0	0	0	0	0	2	0	0
III	14	8	20	0	0	0	-	0	-	0	0	0
IV	33	0	0	-	-	-	-	-	0	0	0	0

TABLE 17 - % FREQUENCY OF VARIOUS AMOUNTS OF RAIN WITHOUT THUNDER (mm).

KEW

	Nil	0.1-0.9	1.0-1.9	2.0-2.9	3.0-3.9	4.0-4.9	5.0 and over
IA	43	18	9	8	6	4	11
IB	44	19	10	7	4	3	13
IC	50	7	11	14	4	4	11
ID	56	18	9	4	2	3	9
II	75	14	3	3	1	1	2
III	43	19	7	7	3	4	18
IV	38	15	7	13	3	5	18

VALENTIA

	Nil	0.1-0.9	1.0-1.9	2.0-2.9	3.0-3.9	4.0-4.9	5.0-5.9	6.0-6.9	7.0-7.9	8.0-8.9	9.0-9.9	10.0 and over
IA	11	18	14	10	9	9	6	2	3	3	3	13
IB	16	21	10	11	7	5	4	4	4	3	1	14
IC	68	18	5	-	-	-	-	5	5	-	-	-
ID	16	19	8	10	8	4	5	3	2	3	3	17
II	61	21	7	2	0	2	2	1	1	1	0	3
III	26	16	8	11	7	5	4	3	3	2	2	15
IV	12	11	9	8	-	12	5	5	6	7	4	22

TABLE 18 - THE ARITHMETIC MEANS OF SPEEDS OF TRAVEL OF DEPRESSIONS (FROM ALL DIRECTIONS) IN MILES PER 12 HOURS. The figures in brackets give the number of observations used in each case.

	IA	IB	IC	ID	II	III	IV
Winter	320(102)	300(94)	266(14)	356(69)	387(45)	303(61)	329(28)
Spring	327(75)	298(177)	289(35)	369(30)	327(85)	310(71)	314(14)
Summer	284(168)	262(173)	310(22)	290(83)	283(46)	287(6)	-
Autumn	314(144)	308(73)	260(19)	344(97)	351(68)	289(19)	307(58)
Year	311(489)	292(517)	281(90)	340(279)	337(244)	297(157)	317(100)

Mean of all types 311

TABLE 19 - PERCENTAGE DIRECTION FREQUENCIES OF MOVEMENT OF DEPRESSIONS.

IA

	Win.	Spr.	Sum.	Aut.
N	5	6	6	5
NW	25	19	13	24
W	40	20	29	30
SW	21	34	26	23
S	7	11	12	12
SE		1	5	2
E		1		1
NE	1	1	2	1
Indefinite	1	5	7	2

IB

	Win.	Spr.	Sum.	Aut.
N	10	7	4	3
NW	21	17	10	13
W	26	23	22	25
SW	20	23	31	42
S	15	16	18	6
SE	6	5	5	6
E	1	4	7	3
NE	1	3	3	1
Indefinite		1	2	

IC

	Win.	Spr.	Sum.	Aut.
N	7	11	14	6
NW	43	17	9	41
W	21	22	14	35
SW		22	32	12
S		6	18	
SE	7	6	9	
E		8		
NE	21	6	5	6
Indefinite		3		

ID

	Win.	Spr.	Sum.	Aut.
N	4	3	4	5
NW	10	20	13	18
W	37	13	30	32
SW	44	50	30	33
S	3	3	8	8
SE		3	7	2
E	1	3	6	
NE		3	1	
Indefinite				1

II

	Win.	Spr.	Sum.	Aut.
N	20	8	13	6
NW	27	38	9	26
W	20	19	37	19
SW	16	20	37	31
S	4	2	2	9
SE		5		
E	4	4		1
NE	9	5	2	6
Indefinite				3

III

	Win.	Spr.	Sum.	Aut.
N	5	6		
NW	15	15		21
W	18	15		11
SW	34	37		22
S	13	11		21
SE	2	1		
E	5	8		11
NE	8	6		5
Indefinite				

IV

	Win.	Spr.	Sum.	Aut.
N		7		10
NW	7	29		10
W	43	14		20
SW	36	36		36
S	11			18
SE	4	7		
E				4
NE		7		
Indefinite				

TABLE 20 - % FREQUENCY OF DIFFERENT SPEEDS OF DEPRESSIONS IN MILES PER 12 HOURS.

	1-99	100-199	200-299	300-399	400-499	500-599	600-699	700-
Winter	1	14	28	26	15	9	4	2
Spring	1	15	32	26	15	7	4	1
Summer	-	20	35	27	10	4	2	1
Autumn	-	13	29	28	17	8	4	1

TABLE 21 - VECTOR MEANS OF DEPRESSION SPEED AND DIRECTION

Season	Type	Speed (miles per 12 hours)	from Direction
Winter	IA	247	W. by N.
	IB	161	W.N.W.
	IC	171	N.W. by N.
	ID	297	W. by S.
	II	214	N.W. by W.
	III	148	W.S.W.
	IV	255	W. by S.
Spring	IA	200	W.
	IB	140	W. by S.
	IC	121	W.N.W.
	ID	257	W.S.W.
	II	200	W.N.W.
	III	151	W. by S.
	IV	179	W. by N.
Summer	IA	179	W.
	IB	146	W.S.W.
	IC	124	W.S.W.
	ID	173	W. by S.
	II	207	W.
	III	150	S.E.
	IV	-	-
Autumn	IA	211	W.
	IB	195	W.S.W.
	IC	195	N.W. by W.
	ID	236	W.
	II	206	W. by N.
	III	130	W. by S.
	IV	203	W.S.W.

TABLE 22 - CLASSIFICATION OF WEATHER TYPES. March 1941 - December 1941.

(Continuation of Table 1)

1941	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1	ID	IB	III	ID	ID	IC	ID	ID	II	III
2	ID	IA	IB	ID	ID	IA	ID	II	II	II
3	IB	IB	ID	II	ID	IA	II	II	II	ID
4	IB	IB	ID	II	IA	IB	II	II	II	ID
5	IB	III	ID	IC	IA	IA	II	II	II	ID
6	IC	III	IB	IB	ID	IA	III	II	IB	ID
7	IB	III	IB	IB	ID	IA	III	II	IB	ID
8	IC	II	IB	IB	IA	IA	II	II	III	IA
9	IB	II	II	IB	IA	IA	II	III	IV	IA
10	II	II	II	IB	IA	IA	II	IB	IA	IA
11	IC	ID	II	IB	IA	IA	II	IA	III	IA
12	II	ID	II	IB	IA	IA	II	ID	III	IA
13	II	ID	II	IA	IB	IA	II	ID	III	IA
14	II	ID	III	IA	IB	IB	II	ID	III	IA
15	II	IA	II	IA	IA	IB	II	IA	II	IA
16	II	IA	II	IA	IA	IB	II	IA	IV	IA
17	II	IA	IA	ID	IA	IA	II	ID	IB	IC
18	II	IA	IA	ID	ID	IB	II	IA	ID	II
19	II	IB	IA	ID	ID	IB	II	IA	IV	II
20	II	IA	IA	ID	IA	IB	II	ID	ID	II
21	IA	IB	IA	II	IA	IA	II	ID	IV	II
22	IB	IB	IA	ID	IA	IA	II	II	ID	II
23	IB	III	IA	IB	IA	IA	II	II	IB	II
24	IB	III	IA	IB	IA	IB	II	II	ID	II
25	IB	IB	IA	IA	IA	IA	IV	II	ID	II
26	IB	IB	IA	IA	IA	IA	ID	II	ID	II
27	IB	IB	IB	IA	IA	IA	IB	II	ID	II
28	IB	IB	IB	ID	IA	IA	IB	II	ID	II
29	IC	III	IB	ID	IA	IB	IB	II	II	II
30	IB	III	IB	ID	IB	IA	IA	II	III	II
31	IB	-	IB	-	IC	ID	-	II	-	II

1942	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
1	ID	ID	IA	IA	III	IB	ID	IA	IA	IA	IA	III
2	ID	IA	IB	IA	IB	II	ID	IB	IA	ID	IB	IA
3	ID	ID	II	IB	IB	II	IA	IB	IA	ID	IB	ID
4	ID	IA	II	III	II	II	IA	IB	IA	IB	IB	IV
5	II	II	III	III	II	II	IA	ID	ID	IV	IA	ID
6	II	II	III	III	IV	II	IA	ID	IA	ID	IB	IA
7	II	II	III	IA	IB	II	IA	ID	ID	ID	III	ID
8	IC	II	IB	IB	IB	II	IA	ID	ID	IA	IA	ID
9	IC	II	IB	IB	IB	II	IC	IA	II	IA	ID	IV
10	ID	II	III	ID	IB	II	IC	IA	II	IA	ID	IV
11	ID	II	III	IV	IB	III	IA	IA	II	IA	ID	ID
12	ID	IA	III	III	III	III	IA	IA	II	ID	ID	IV
13	IA	IA	III	II	III	III	IA	IA	II	ID	ID	IV
14	IB	II	IV	II	IB	III	IA	IA	II	IV	ID	IA
15	IB	II	IV	II	IB	IA	IA	IA	ID	ID	II	IB
16	IB	II	IB	II	IB	IA	IA	ID	IB	IA	II	IB
17	IB	II	IA	II	IB	IA	IC	ID	II	IA	II	IB
18	ID	II	IB	IB	IB	ID	IC	IA	IC	IB	II	IA
19	ID	II	IB	IB	IB	ID	IA	IA	ID	IB	II	IA
20	ID	III	III	IA	ID	II	IA	IB	IA	IB	II	IA
21	ID	III	II	IA	IA	ID	IA	IB	IA	ID	II	ID
22	ID	III	II	IA	IA	ID	IA	ID	IA	ID	II	ID
23	ID	III	II	II	IA	ID	IA	IA	IA	ID	II	ID
24	ID	III	II	II	IA	II	IA	IB	IA	IB	II	ID
25	IA	III	II	II	ID	II	IA	IB	IA	IB	II	ID
26	IA	III	II	III	IA	II	IA	IA	IA	IB	II	ID
27	IA	III	III	III	IB	ID	IA	IA	IA	IA	II	ID
28	IA	III	III	III	IB	IA	IA	IA	IA	III	II	ID
29	IA	-	II	III	IB	IA	ID	IC	IB	III	III	ID
30	IA	-	IA	III	IB	IA	ID	IB	IB	IC	III	ID
31	IA	-	ID	-	IB	-	ID	IB	-	IC	-	ID

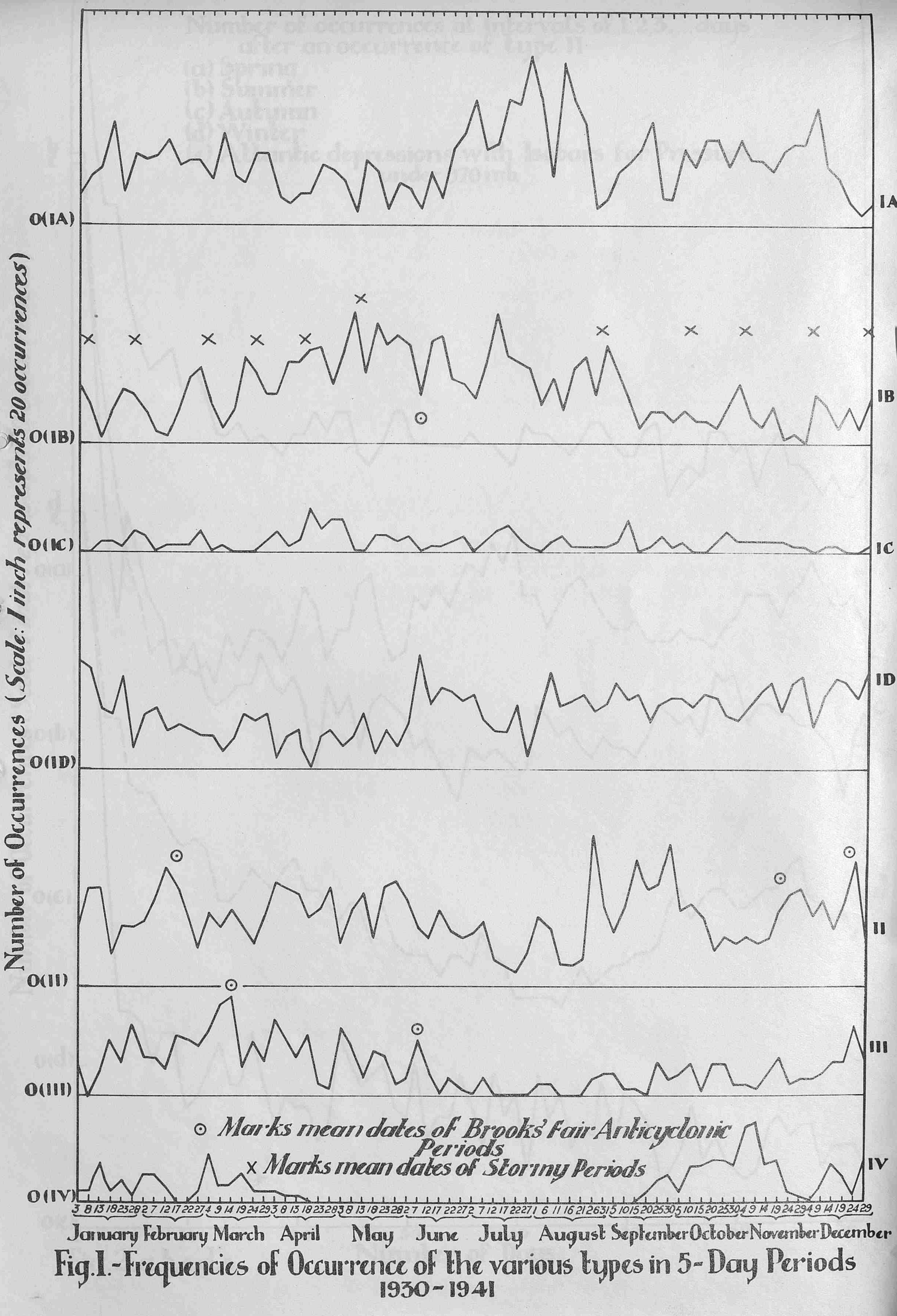


Fig.1.-Frequencies of Occurrence of the various types in 5-Day Periods
1930-1941

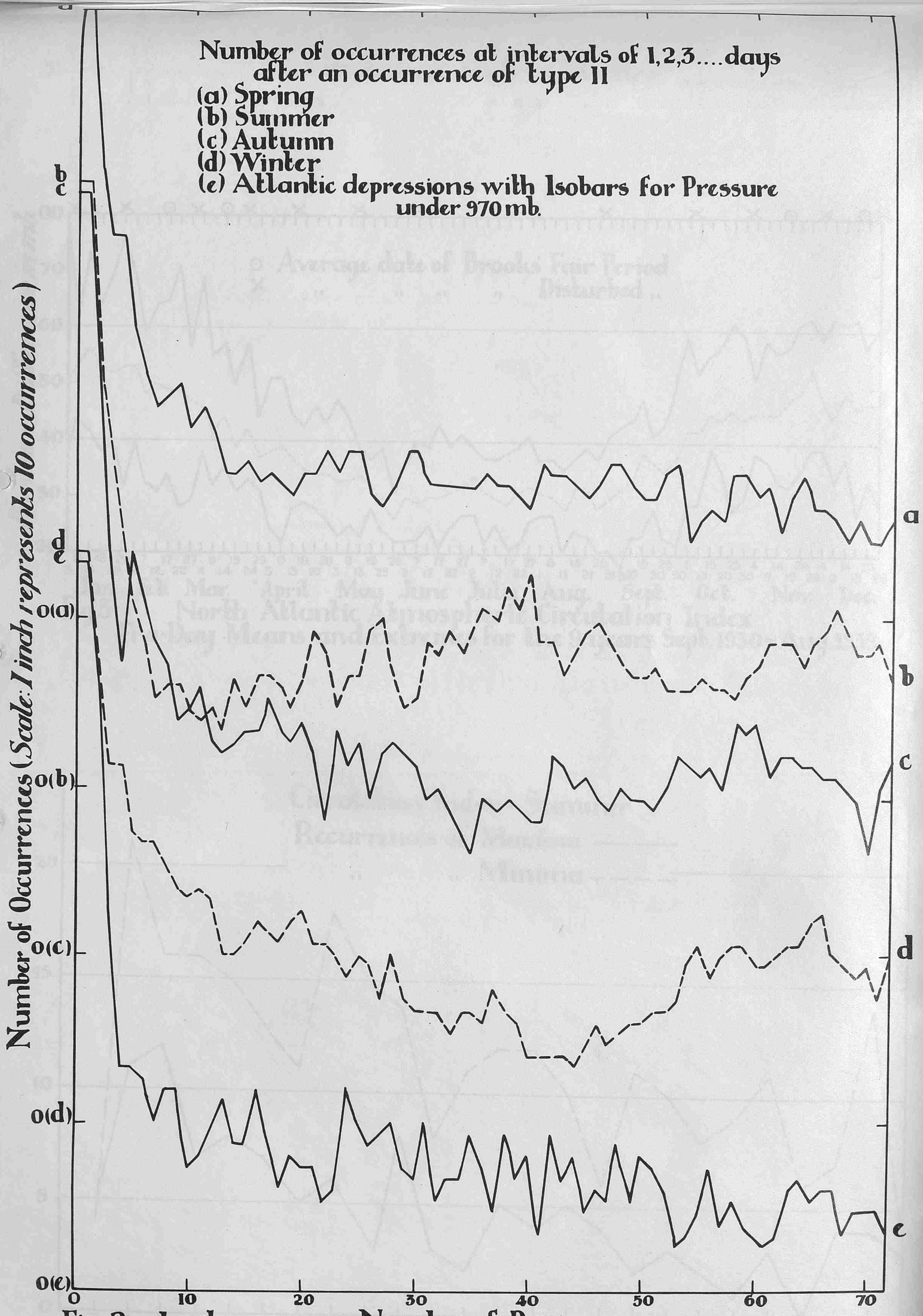

Number of occurrences at intervals of 1,2,3....days after an occurrence of type II
(a) Spring
(b) Summer
(c) Autumn
(d) Winter
(e) Atlantic depressions with Isobars for Pressure under 970 mb.
Number of Occurrences (Scale: 1 inch represents 10 occurrences)
b
c
d
e
0(a)
0(b)
0(c)
0(d)
0(e)
a
b
c
d
e
0
10
20
30
40
50
60
70
Number of Days

Fig. 2, a, b, c, d, e.

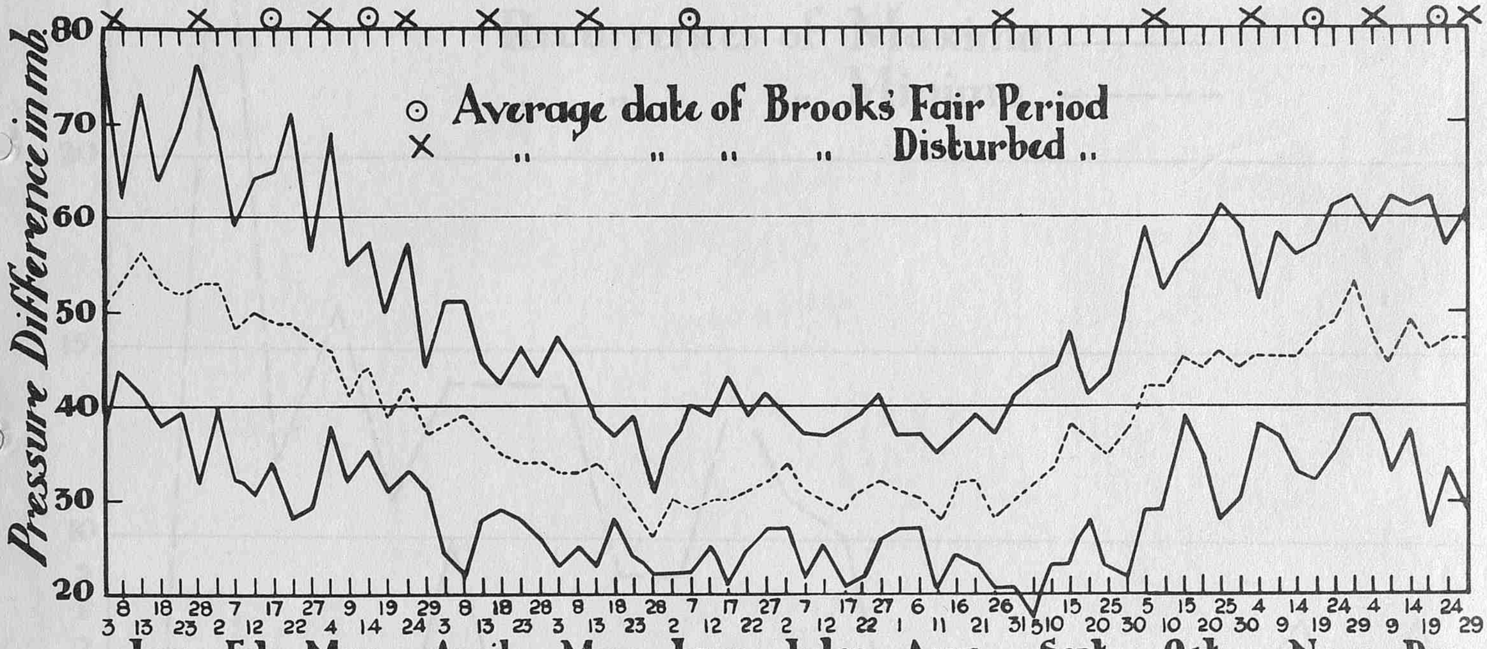


Fig.3. North Atlantic Atmospheric Circulation Index
Five-Day Means and Extremes for the 9 years Sept. 1930~Aug. 1939

Fig. 4(a) Number of Days after First Maximum or Minimum

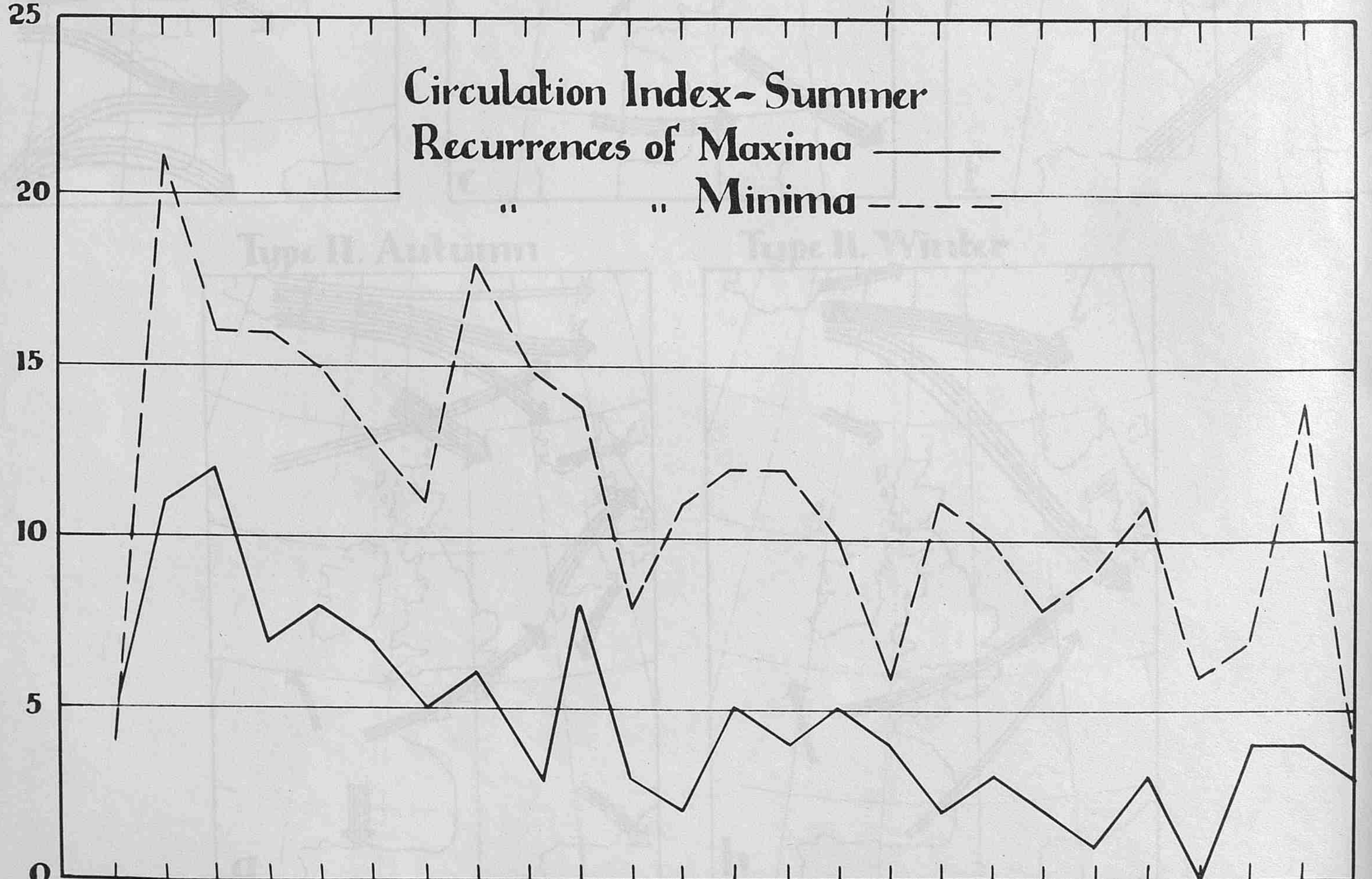


Fig. 4(b) Number of Days after first Maximum or Minimum

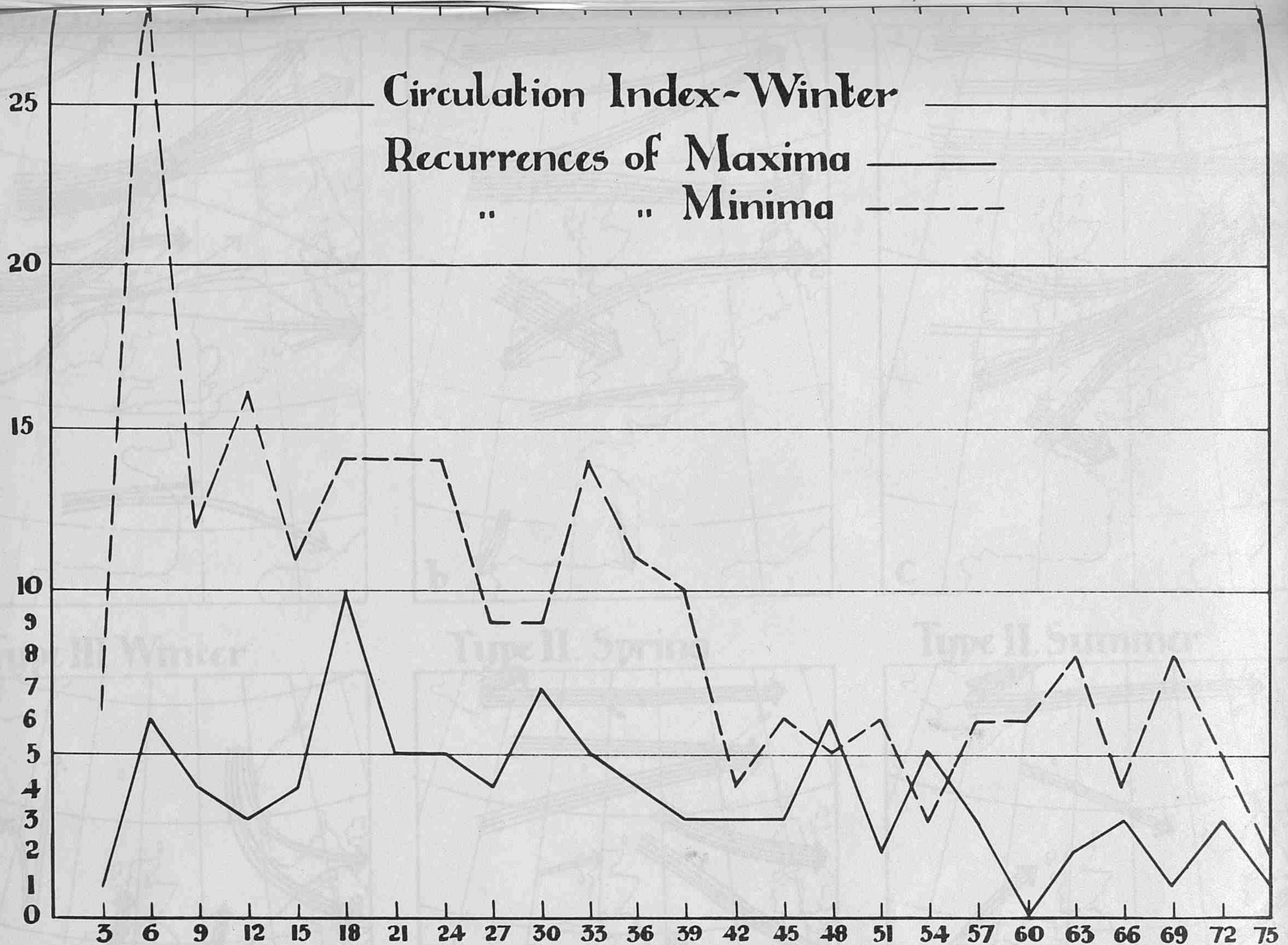


Fig. 4(a) Number of Days after first Maximum or Minimum

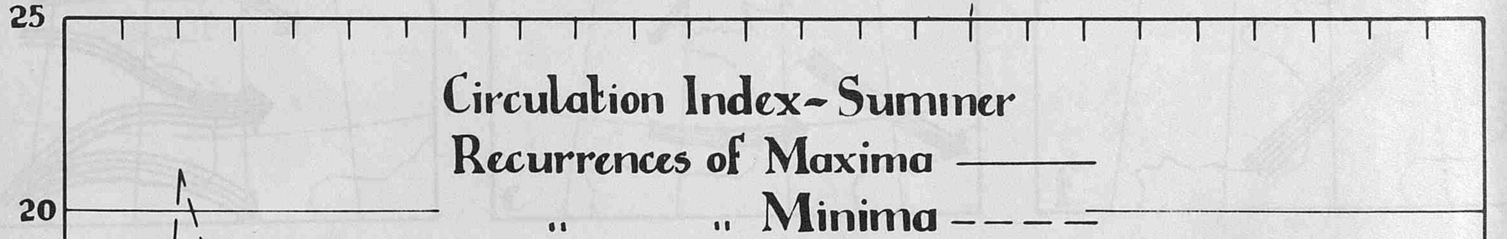
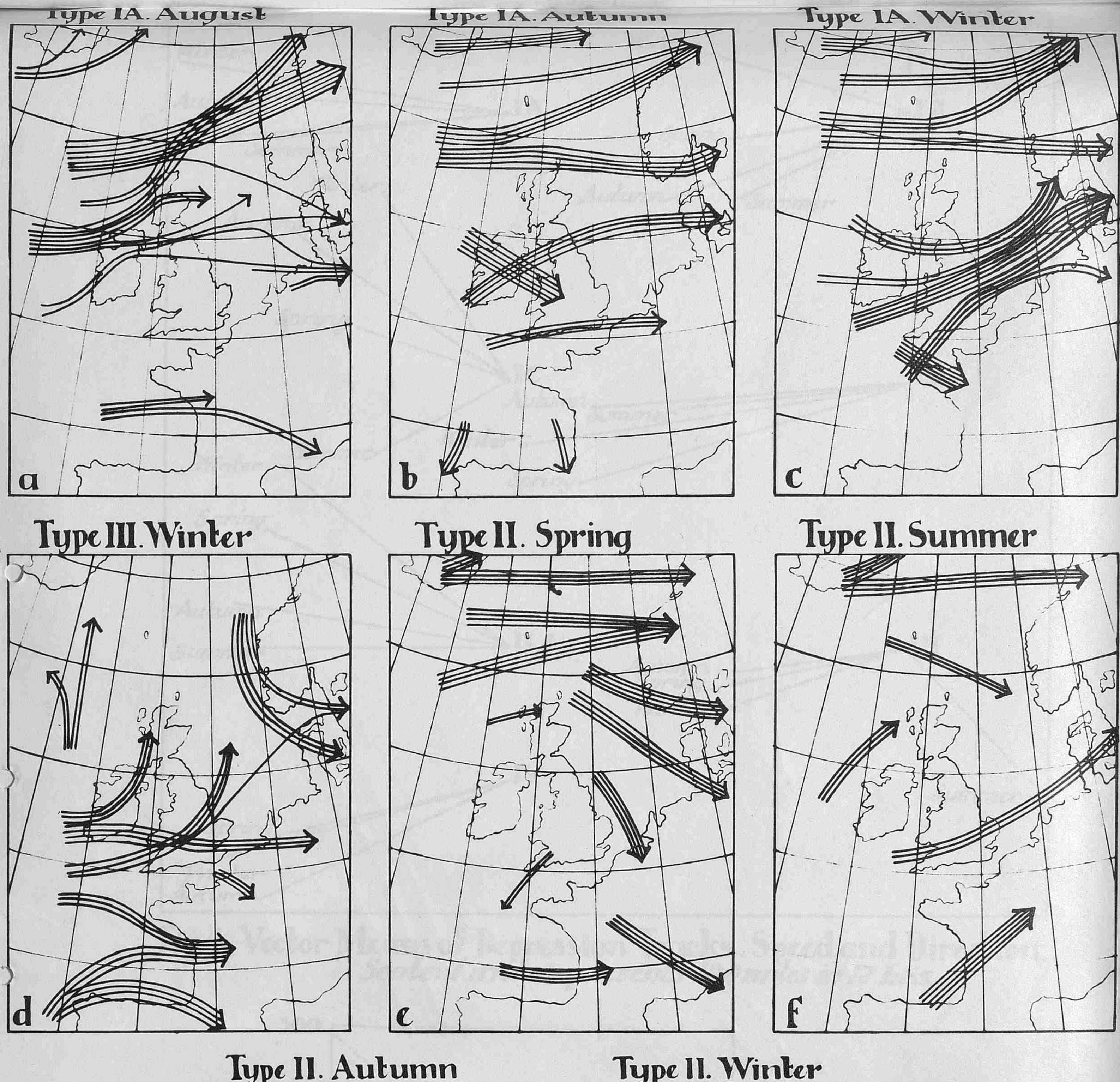


Fig. 4(b) Number of Days after first Maximum or Minimum



Type IA. August

Type IA. Autumn

Type IA. Winter

Type III. Winter

Type II. Spring

Type II. Summer

Type II. Autumn

Type II. Winter

Fig. 5. Depression Tracks 1930-1941

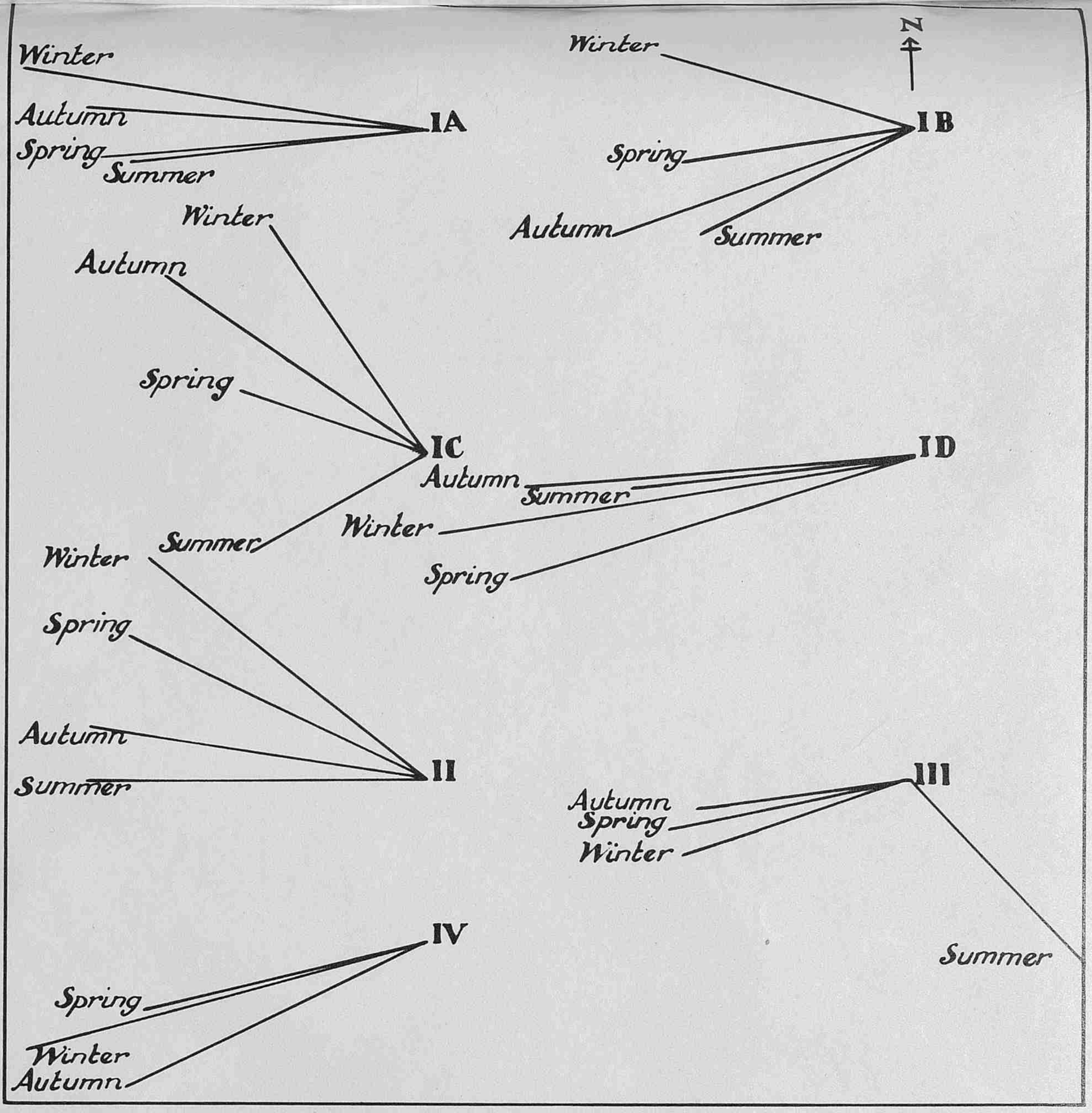

N
Winter
Autumn
Spring
Summer
IA
Winter
Spring
Autumn
Summer
IB
Winter
Autumn
Spring
IC
Autumn
Summer
IC
Autumn
Summer
Winter
Spring
ID
Winter
Spring
Autumn
Summer
II
Autumn
Spring
Winter
III
Summer
IV
Spring
Winter
Autumn

Fig. 6. Vector Means of Depression Tracks. Speed and Direction.
Scale: 1 inch represents 100 miles in 12 hrs.

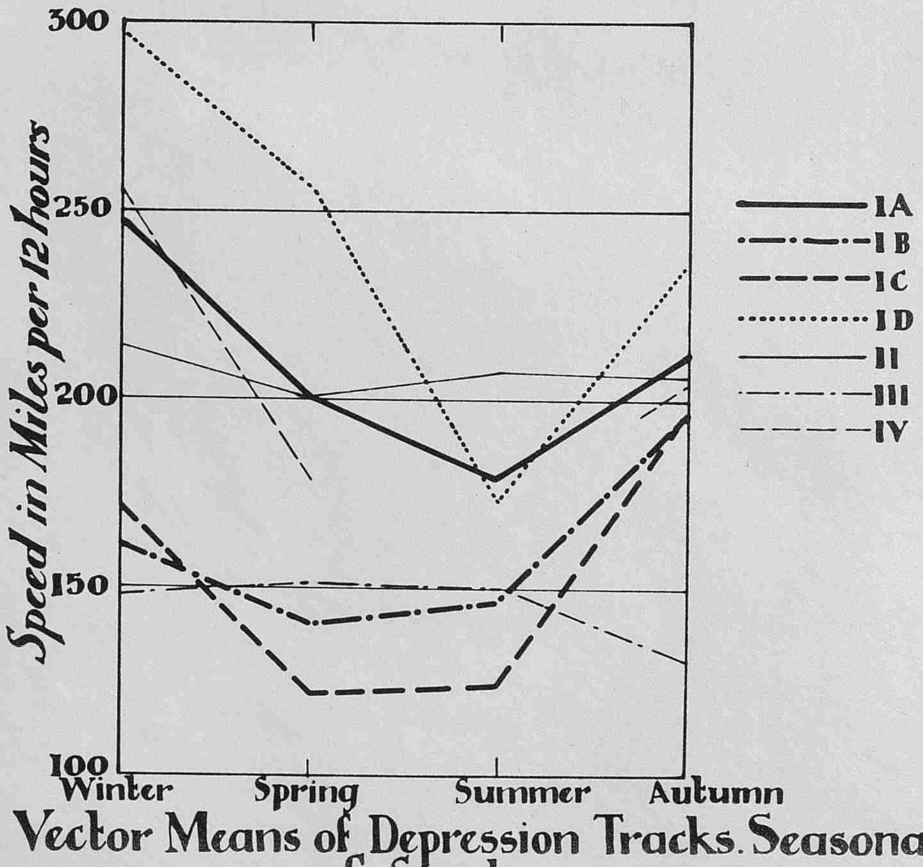


Fig. 7. Vector Means of Depression Tracks. Seasonal Variation of Speed.

leave a permanent pit after they evaporate.

Laboratory experiments with droplets of substances such as dibutyl pathalate which can be sized accurately by other means show that at a speed of impact of 10 m.p.h. the diameter of the hole in the MgO is the same as that of the original drop. Viewed by transmitted light the holes are seen as bright white spots on a darker background. At aircraft speeds drops as small as $3\text{-}4\mu$ leave a visible trace.

The MgO method has advantages over the absolute method in that:-

(1) It is much simpler as nothing has to be done to the plate after the sample is obtained.

(2) The craters remain without change for a long period of time (2 - 3 days).

Against these:-

(1) The layer is very soft and fragile, and therefore it may be blown away when the speed is great.

(2) The visibility of drops of colourless liquids is not good (at low speeds) when the diameters are below 10 microns.

(3) The method is of no value below $3\text{-}4\mu$ (The absolute method can show drops below 2μ diameter).

To compare the results by the two methods the glass plates are coated half with MgO and half with the oil-vaseline mixture.

V. Drop Counting Technique

In the method used at Porton droplets are counted against a Patterson-Cawood globe and circle comparator graticule, traverses being taken across a mechanical stage. For a good count about 300 drops should be taken. The number percentage in the ranges between the mean diameters corresponding to adjacent pairs of circles are then worked out and a rough curve of per cent. by number up to given diameter is plotted against that diameter on a logarithmic probability paper. From a smoothed curve through the plotted points a new number distribution is taken and converted into a table of total number, N, in the size range tabulated against size range. The mean diameter D of each size range is cubed and then multiplied by the number of drops in the range to give a table of ND^3 which, when reduced to a percentage, gives a mass or a volume distribution. The total mass in one hundred drops is equal to

$$M = \frac{\pi\rho}{6}\Sigma ND^3,$$ where ρ is the density in g/cc.

The concentration in g/m^3 can be calculated, and the mean volume diameter can also be calculated from

$$\overline{Dv} = \sqrt[3]{\frac{ND^3}{100}}$$

VI. The Air Speed through the Aircraft-Impactor

In such a small apparatus it is difficult to obtain trustworthy readings of air speed.

Two sets of experiments were made. For the first, an old pitot-static tube - external diameter just under $\frac{1}{4}$ inch, and spacing between static and pitot holes 0.65 inches - was used. Inside a slot which measures only 0.6 x 1.4 inches this distance between pitot and static holes is obviously a serious disadvantage. For the second set of experiments a specially fine pitot tube was constructed, using two stainless steel tubes each of 0.9 mm. diameter and 35 cm. length soldered together. One advantage of the second type is that it does not disturb the flow of air.

......

Measurements with both types of pitot were made at speeds of about 70 m.p.h. and 50 m.p.h. The results of the calibrations are shown in Figs.3(a) and 3(b) respectively. From the latter, in which the fine pitot was used, it is seen that the velocity of the air sample in the slot is approximately equal to the velocity of the apparatus through the ambient air.

VII. Observations made in Natural Clouds

Observations in natural clouds have been made over Salisbury Plain, using a Hampden bomber from Boscombe Down aerodrome.

Consecutive samples in the same cloud formations at different points both in horizontal and vertical positions were taken during every flight.

The accuracy with which the drop diameter can be measured is limited by the resolving power of the microscope and the grain of the film. The grain size of the KODAK plates used is so small that it does not interfere with the definition of particles at least down to 1μ. It was possible during these experiments to see drops of about 1μ in diameter.

No flattening effect of the drops was seen.

The curves which represent the drop size distributions or the size frequency for a cloud are very similar to a Gaussian or normal distribution curve (Maxwell curve).

The most obvious and characteristic feature common to all of the curves is that there is a single and very strongly marked maximum. In none of the samples so far examined was there more than one peak in the size distribution curves.

The distribution curves evidently become asymptotic to the axis for larger particle sizes and a definite and very sharp lower limit of drop size exists.

The size of water particles varies sometimes from one part of a cloud to another. This is true at least in the case of Cumulus clouds. There are types of cloud in which the particle size remains sensibly constant, e.g. Altostratus.

The locations of the maxima of the distribution curves vary not only for different clouds but within the same cloud.

The absence of secondary maxima is an indication that there has been no appreciable combining of the drops.

In the interior of clouds there are larger drop sizes than in the boundary layers where the droplets must be in equilibrium with saturated ambient air.

The range of size is as follows: the diameter of the smallest drops was less than 1μ, and that of the largest about 100μ. The water content in clouds varies between 0.12 g/m^3 and 0.41 g/m^3.

Four microphotographs of cloud drops embedded both in the grease film and in a magnesium oxide layer are shown in Figs.4, 5, 6, 7. As showing typical size distribution curves for a cloud the frequency diagrams for 10 flights are given in Figs.8 to 17. A summary of the data is shown in Table I.

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TABLE I

Size of drops and water contents in different clouds

Date of flight	Genus	Diameter of the drops			Concentration
		Min.	Mean	Max.	
22.1.43	Stratocumulus	2 μ	24.2 μ	100 μ	0.16 g/m³
27.1.43	Stratocumulus	3 "	14.4 "	45 "	0.12 "
2.2.43	Cumulus	9 "	14.7 "	50 "	0.13 "
10.2.43	Stratocumulus	3 "	13.1 "	37.5	0.12 "
12.2.43	Stratocumulus	1.5"	13.3 "	37.5	0.15 "
17.2.43	Altostratus	1 "	10.6 "	60 "	0.13 "
8.4.43	Cumulus	1.4"	8.8 "	21 "	0.12 "
(+) 11.5.43	Cumulus	3 "	10.2 "	18 "	0.41 "
(+) 13.5.43	Stratocumulus	4 "	9.0 "	16 "	0.38 "
(+) 1.6.43	Cumulus	6 "	12.6 "	18 "	0.20 "

(+) New Aircraft Impactor has been used.

VIII. Acknowledgments

The investigation was carried out at the Physics Department, Experimental Station, Porton and all flights were made from Boscombe Down aerodrome.

Some experiments concerning the behaviour of the drops were carried out at the Physics Department, Imperial College of Science and Technology.

I wish particularly to acknowledge the assistance given by Mr. H. L. Green, Mr. C. N. Davies, Mr. K. R. May, Messrs. Dennis and Prewett, Miss E. Bowen and Mr. Edwards of Porton who have helped in different ways in the design and construction of the Impactor, in the making of calculations and in the experiments and observations.

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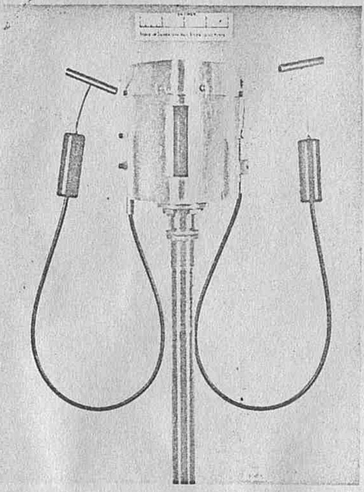


Fig 1a. Front view of Aircraft Impactor showing front shutter open.

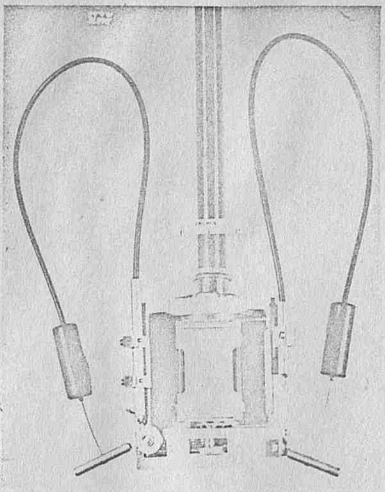


Fig. 1b. Back of the Aircraft Impactor showing plate holder swung round to position for changing plate.

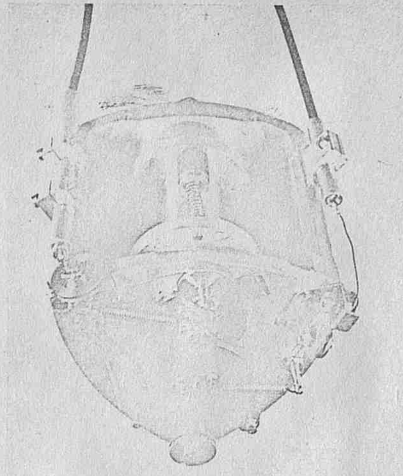


Fig. 2. Aircraft Impactor. View with cover removed to show the actuating mechanism.

Fig. 3(a).

CALIBRATION OF SPEEDS IN AIRCRAFT IMPACTOR

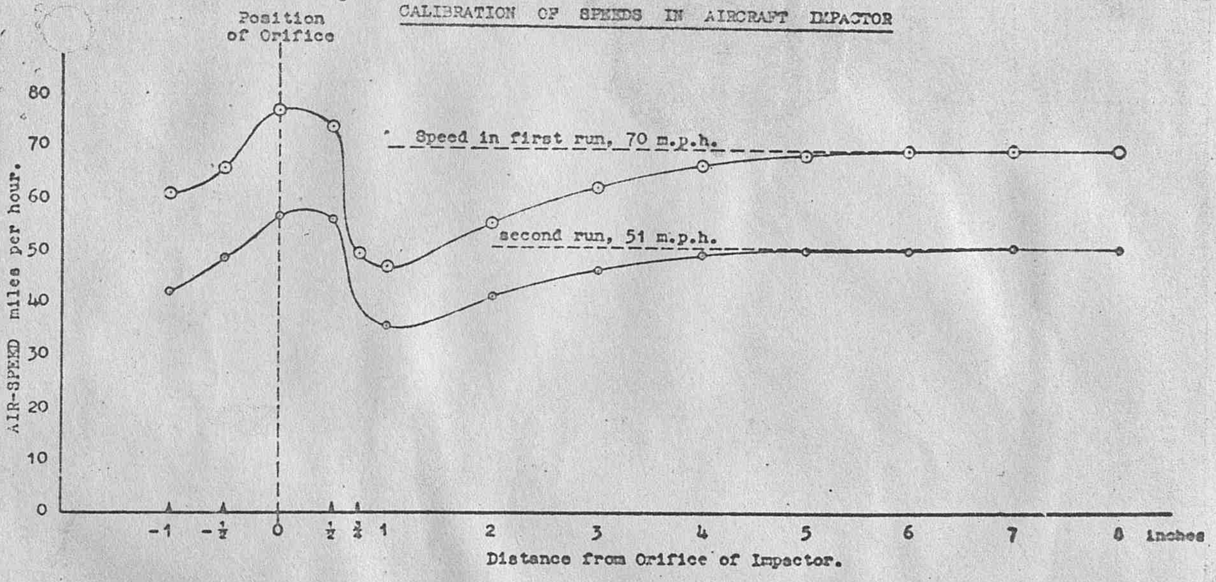
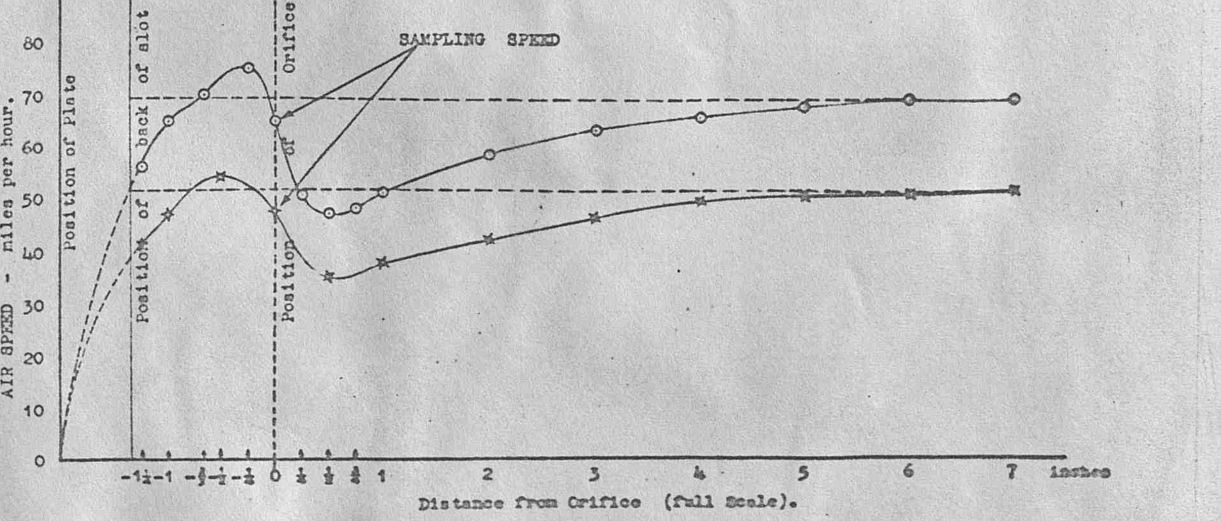


Fig. 3(b).

AIR SPEEDS ALONG CENTRAL AXIS OF AIRCRAFT IMPACTOR (by means of a specially constructed Pitot-tube.)



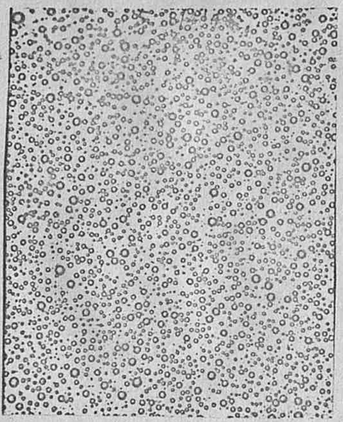


Fig. 4. Cumulus, 4500 ft. X 100
Flight on 8-4-43

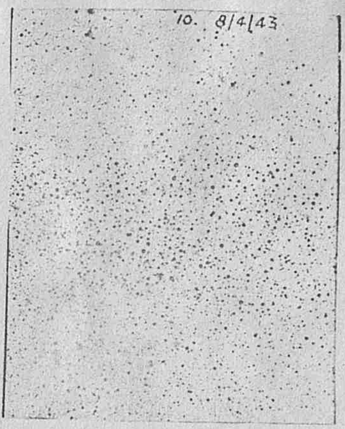


Fig. 5. Cumulus, 4500 ft.
Boundaries of cloud X 100
Flight on 8-4-43

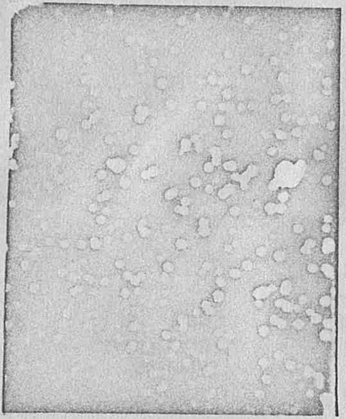


Fig. 6. Altostratus, 8000 ft. X 100
Flight on 17-2-43

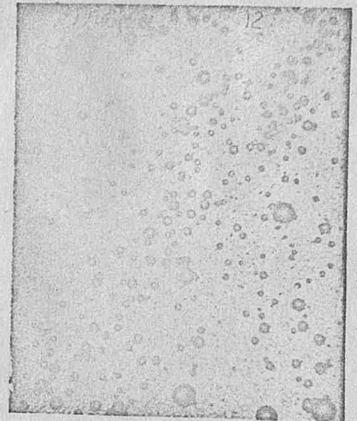


Fig. 7. Altostratus, 8000 ft.
The same plate as Fig. 6 X 100
Flight on 17-2-43

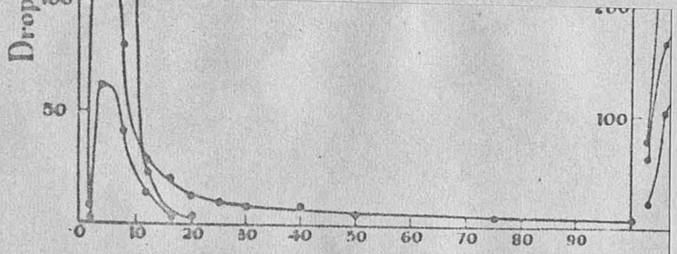


FIG. 8

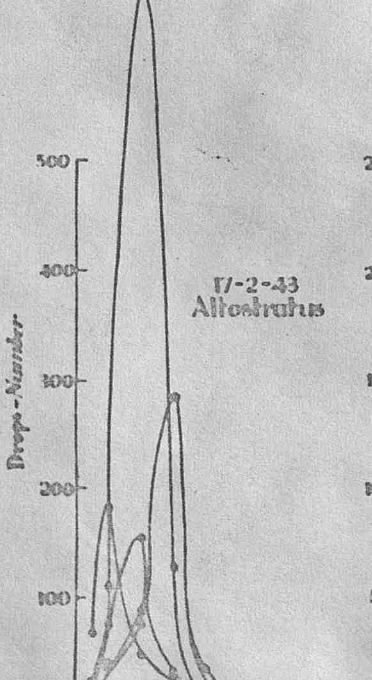


FIG. 13

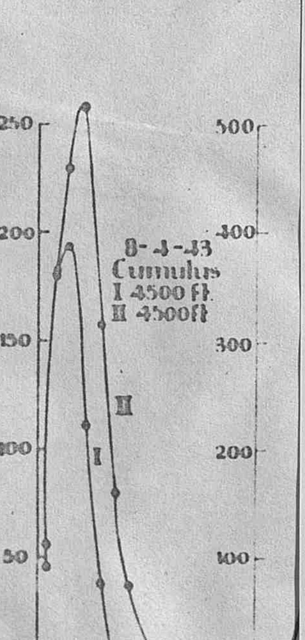


FIG. 14

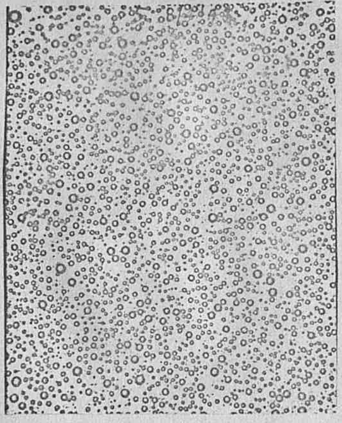


Fig 4. Cumulus 4500 ft X 100

Fig 5. Cumulus [illegible]

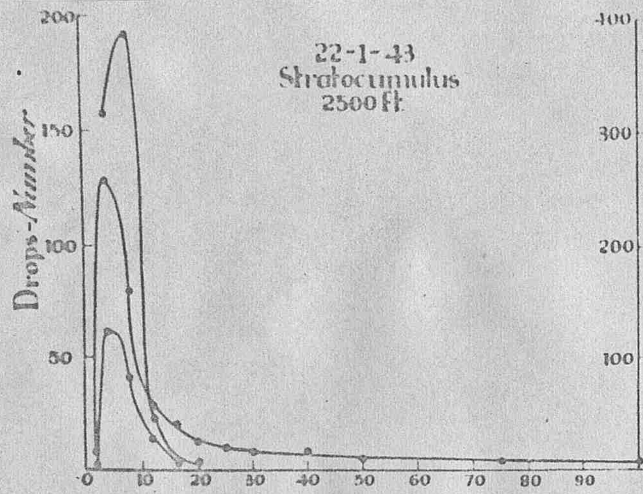


FIG. 8

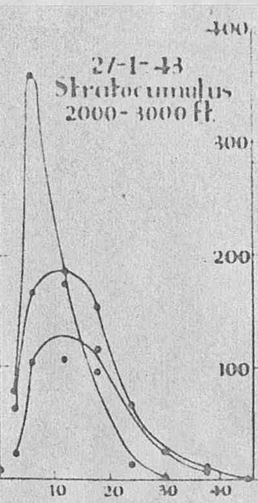


FIG. 9

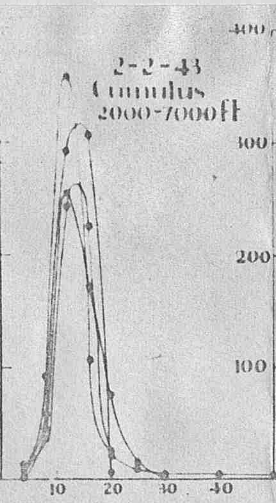


FIG. 10

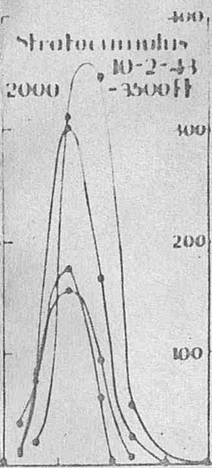


FIG. 11

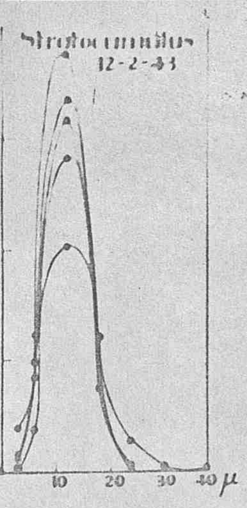


FIG. 12

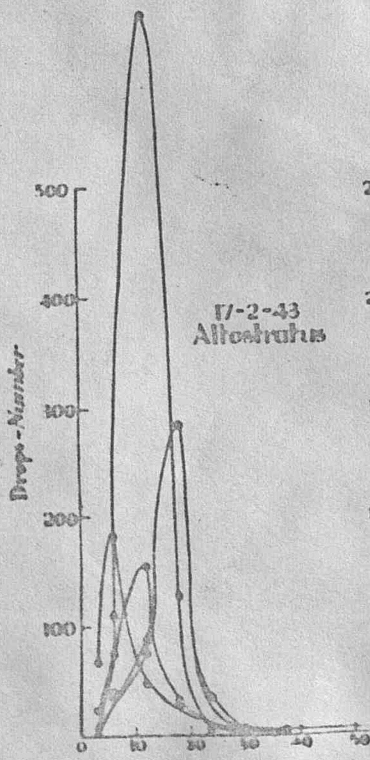


FIG 13

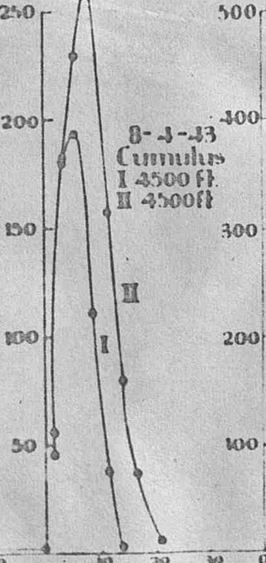


FIG 14

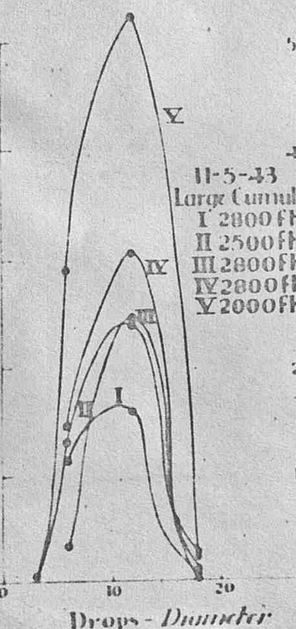


FIG. 15

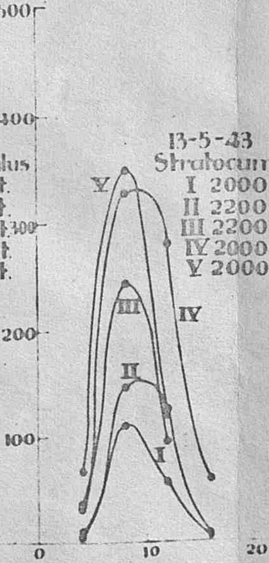


FIG. 16

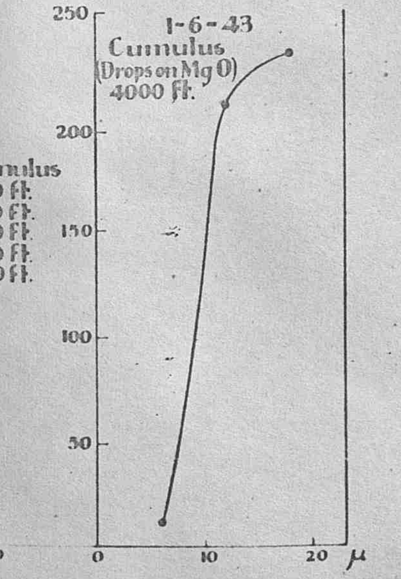


FIG. 17

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An Investigation on the Number and Size Distribution of Water Particles in Natural Clouds

Report by Dr. Jozef Mazur,
F/Lt. Polish Air Force

This report has not yet been published and is to be treated ~~CONFIDENTIALLY~~. Its contents must not be quoted.

I. Introduction

The danger to aircraft of ice formation has given rise to a number of experiments in which the ice formation is produced artificially under controlled conditions of temperature, airspeed, size and number of suspended water particles. The application of the results of such experiments depends on information about the number and size distribution of water particles in natural clouds. In January 1942 I prepared a memorandum summarizing the existing knowledge on this subject and indicating the lines on which further investigation was needed. As a first step it was necessary to devise an apparatus for sampling the water drops, to develop a technique and begin observations.

II. General Description of Apparatus

The apparatus, known as the Aircraft Impactor was conceived on the lines of the Porton Impactor. The latter, designed by Mr. K. R. May, consists of a wood and metal holder for a glass plate. The wooden front is "streamlined" and has a slot through which the natural wind passes, thus playing on the surface of the glass plate clamped at right-angles to the outlet of the slot. The air stream impinging upon the plate escapes round two sides of the plate. The air has to turn through a right angle, and the particles having sufficient kinetic energy are unable to turn the corner and so strike and adhere to the plate. The impactor is used in its vertical position so as to minimise the sedimentation effect. The Impactor method depends upon allowing the drops to impinge at a high velocity upon a plate covered with a suitable film in which the drops are trapped. When the impactor has to be used in cloud the velocity of impact is provided by the motion of the aircraft so that somewhat different problems have to be faced; also the possibility that ice formation might impede the functioning of the moving parts had to be considered.

III. The Aircraft Impactor

The first type of aircraft impactor, made in the Instruments Branch of the Meteorological Office at Stonehouse, retained the shape of the Porton impactor but differed in regard to the shutters. These consisted of two co-axial semi-cylindrical cases, arranged to rotate in opposite directions and each provided with a slot parallel to the axis. By suitable setting of a spring controlling the mechanism the slots could be brought together to give the correct exposure.

After trials of this first model in a wind tunnel and on an aircraft, an improved impactor was made. The body was cut from a solid aluminium block and is very strong, the slot was shortened and the sampling plate reduced to half the size of a quarter plate.

Photographs of the instrument are shown in Figs. 1(a), 1(b) and 2.

Fig. 1(a) shows the front of the instrument with the front shutter open. Behind this shutter there is another shutter whose time of opening can be calibrated by means of a cathode ray oscillograph and a tuning fork. Of the two cables shown one is a "release cable"; when it is pulled the shutters work in the following order:-

(1) The front shutter opens and is held by a stop. When it reaches the stop, the rear calibrated shutter is released.

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(2) The rear (or calibrated) shutter exposes and when the exposure is finished, the front shutter stop is released and the calibrated shutter is held by the recoil check lever. This is to prevent the shutter bouncing back and re-exposing.

(3) The front shutter closes.

When the other cable - the "rewind" or reset cable is pulled the following operations take place:-

(1) The recoil check lever is disengaged.

(2) The calibrated shutter is rewound.

(3) The front cylindrical shutter is rewound.

(4) The rewind plate returns to its normal position when the rewind cable is released.

The plate holder seen in Fig.1(b) consists of an aluminium sheet with pieces cut out in the vertical sides to facilitate changing plates. In the operational position this plate holder is mounted on ball bearings (taken from a German plane shot down over Salisbury Plain) and can be turned easily when plates have to be changed.

IV. The Sampling Plate

The surfacing of the plates used for the sampling of the drops is of the utmost importance. The plates are washed first in concentrated chromic acid, then in distilled water and allowed to dry in a dust-free atmosphere. The receiving surface can then be coated with a uniform film of a substance in which the water particles can be trapped. Two methods of coating have been used - the absolute method and the magnesium oxide method.

In the first method the object is to preserve the drops in their original form by enclosing them in a surfacing substance which must have the following properties:-

(1) A density not very different from that of water.

(2) Adequate viscosity.

(3) Durability and non-solubility in water.

(4) Colour contrast with water drops.

(5) Non-freezing and showing little change of viscosity with temperature.

A successful procedure is to catch the drops on a surface consisting of a first layer of freshly melted vaseline and light mineral oil. As soon as possible this is covered with a cover glass on the underside of which is a layer of castor oil.

Castor oil has a density of 0.959 - 0.968 g/c.c. at 15.5°C, and a refractive index at 20°C of 1.777 - 1.429. This is the reason for the good visibility of water drops in Castor Oil. Castor Oil also possesses the highest viscosity of all known fatty oils. This fact is very important because for an oil of given viscosity the higher the viscosity index the less change there is in viscosity over a given temperature range. It does not dry even when exposed to the atmosphere in the form of a thin layer.

The above absolute method is useful as allowing the droplets to be examined microscopically and counted at leisure.

In the second method MgO is "smoked" on the glass plate by means of a gun burning magnesium ribbon. The droplets drive into the soft layer of MgO and

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A DISTANT READING ELECTRICAL AIR TEMPERATURE THERMOMETER EMPLOYING A BALANCED BRIDGE SUITABLE FOR USE IN AIRCRAFT

File
Instruments

By A. W. BREWER. M.Sc.

1. Summary

The requirements for a distant reading thermometer for meteorological flights are outlined. A resistance thermometer employing a balanced Wheatstone's bridge which satisfies the requirements is described, and the results of flight tests given. The instrument employs a Cambridge Instrument Aircraft electrical thermometer resistance element. The rest of the instrument is specially designed.

2. Requirements of a Thermometer for use on Meteorological flights.

a) Existing aircraft air thermometers are deficient in one or more of the following respects:-

(i) Limited range;

(ii) Contracted scale

(iii) Inaccuracy, other than that due to contracted scale.

(iv) Inapplicability to monoplane or Single seater aircraft.

b) The requirements which an aircraft air thermometer should satisfy are:-

(i) Distant reading, - preferably without difficult threading operations, as for instance is required in the mercury in steel thermometer, in which a large bulb must be threaded through the aircraft, or perhaps through an airtight skin of a pressure cabin.

(ii) Range - -100°F. to +100°F.

(iii) Accuracy, ±1°F. This requirement carries with it the need of an adequate scale length and freedom from errors due to changes in the temperature of parts of the apparatus other than the thermometer element and freedom from any effects due to air pressure variation.

(iv) Small indicator to permit installation in a Spitfire or similar cockpit.

(v) Freedom from trouble due to vibration.

(vi) Rapidity of response. The ideal is to have the response sufficiently rapid to permit taking readings during a climb without level flight. In any case it is desirable that the required period of level flight should not exceed about 30 secs.

c) The above requirements can be more easily met in a multiseater aircraft in which an observer can give individual attention to the observation and in which ample space to accommodate the instrument is available. If therefore the requirements can be met for the more difficult case of a single-seater aircraft, the instrument will automatically

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satisfy all aircraft needs. We therefore take as a criterion of scale length the discrimination which a pilot could reasonably exercise in a reading at say 40,000 ft. This discrimination is about 0.5 mm. which if made to correspond to 0.5°F., the required accuracy being then readily obtained, would necessitate a scale length of 20 cm. (8 ins.) for the required range of 200°F. A scale length of 20 cm. on a circular dial implies a dial diameter not less than 6.4 cm. (2.5 in.) and more in proportion to the fraction of circumference unused for the scale. A dial of smaller diameter would only be practicable if the instrument were of the multirange type.

d) It was felt that a balanced bridge resistance thermometer was the most promising method of satisfying these requirements and the instrument described here was developed. All the parts are rugged and show no evidence of trouble if subject to normal vibration tests.

3. The sensitive element.

a) The element of the Cambridge aircraft electrical thermometer is a platinum resistance thermometer with a fundamental interval of 40 ohms. It is a standard article of rapid response and is reasonably suited to the present purposes. The development of the balanced bridge indicator was combined with the work of selecting a suitable mounting for this element, as has been described in M.R.P.113

b) The Cambridge "bulb" is not of ideal design. It is rather bulky and does not fit snugly on to modern aircraft; nor is it specially designed to avoid icing or becoming wet in cloud or rain. These defects are not of special concern here as the main problem is that of the indicator which depends on the fundamental interval of the thermometer only. The bridge balancing devices and scale are the essential features of the instrument now being described. A more suitable bulb may be developed later and used instead of the Cambridge bulb.

4. The Balanced Bridge.

a) The Bridge circuit employed is a normal arrangement using compensating leads on the three wire system. A diagram of the circuit is shown in Fig.1

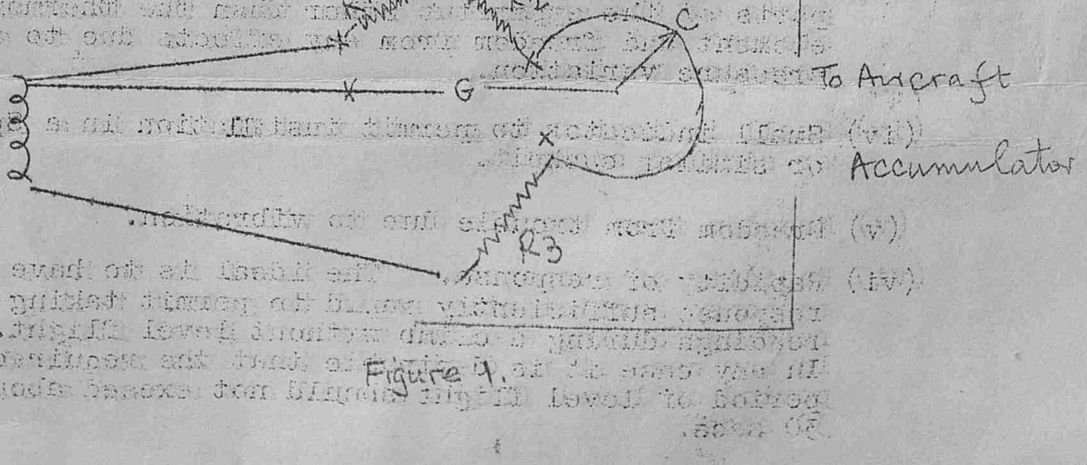

Figure 1.

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b) T is the thermometer and R_1 is a fixed balancing resistance. The other side of the bridge consists of two equal resistances R_2 and R_3 and a slide wire XX which is about 320° of a single turn of suitable wire stretched round the periphery of a circular disc. The galvanometer is connected to a sliding contact C and the bridge is balanced by movement of the contact to obtain zero galvanometer current. If T_{-100} is the resistance of the thermometer element at -100°F. and T_{+100} is the resistance at +100°F. and if W is the total slide wire resistance, we have from the ordinary Wheatstone's Bridge Formula -

$$T_{-100} = R_1 \frac{R_2}{R_3 + W} \quad \text{and} \quad T_{+100} = R_1 \frac{R_2 + W}{R_3}$$

because the sliding contact moves from one end of the slide wire to the other over the range -100°F. to +100°F.

By making $R_2 = R_3$ so that we then obtain

$$R_1^2 = (T_{-100}) \times (T_{+100}) \text{ or } R_1 = \sqrt{(T_{-100}) \times (T_{+100})} =$$

Geometrical mean of T_{-100} and T_{+100}

also R_2 and R_3 must satisfy the condition

$$\frac{T_{+100}}{T_{-100}} = \left(\frac{R_2 + W}{R_2}\right)^2 \quad \text{or} \quad \frac{W}{R_2} = \sqrt{\frac{T_{+100}}{T_{-100}}} - 1.$$

In the bridge described here R_1 = 99 w approx. W = 4.8 ohms. approx. and R_1 and R_2 are 18 ohms approx. R_1, R_2, and R_3 are of constantan and the slide wire is hard drawn stay brite steel.

5. Compensation of Lead Resistance

a) In the Cambridge electrical thermometer referred to in 2a), a rheostat is incorporated in the instrument to take care of differences in the lead resistance between different installations. This has disadvantages in that the rheostat is another item to go wrong and also that changes in lead resistance due to changes in their temperature was not cared for. A compensating system was therefore considered, though perfect compensation can only be attained in a unit ratio bridge. The bridge is only a unit ratio bridge at temperatures in the middle of the range but for the temperature range with which we are concerned, the ratio always lies within the range 0.8 to 12.5. Thus, though perfect compensation could only be attained at certain temperatures 80% compensation would be the worst condition obtained; this would be at each end of the temperature scale.

b) A three wire compensating system was used as it is only a very slight increase in complexity over the non-compensating system, and also, compared with the common 4 wire system, is only half as sensitive to departures of the bridge from unit ratio. The three wire compensation provides a wire in each side of the bridge between the indicator and the distant thermometer element, the third wire carries the galvanometer current. The bridge is normally calibrated with 10 or 15 feet of cable between the element and the indicator, which is about the length used in a Spitfire installation, but if in a particular installation a different length of cable were used, this would have no effect at the centre of the range but would cause a small error at the extremes of the temperature range. The maximum error is quite negligible; for suppose the instrument to be used with 60 feet more of 4 amp. wire (each cable = 13 x 37 S.W.G. copper) then the length for which it was calibrated, a correction of about + ½°F. would be required at the high temperature end of the scale and -½°F. at the low temperature end. Thus it is only with wire lengths far greater than will normally be encountered that any appreciable error results.

6. The Construction of the Instrument

a) The measurement of temperature by a resistance thermometer in a balanced bridge is very common laboratory practice, but for an aircraft installation a very differently constructed instrument is required.

b) Fig.2 is a perspective drawing of the most compact instrument which has yet been made. The complete instrument is built into a box 3½" x 4½" x 2" deep internal dimension and 4½" x 5" x 2½" deep external dimensions. It can be conveniently accommodated in a Spitfire cockpit if the gunsight is removed.

c) The bridge is balanced by rotating the large knob 1, which drives the sliding contact 2, mounted on the arm 3, through a 9-1 ratio indirect drive 4. The slide wire is stretched round the paxolin disc 5 which is built on to the body of the indirect drive. The balance point is indicated by the galvanometer 6. This is a small 2 pivot edgewise moving coil galvanometer. It has a centre zero and pole pieces specially shaped so that maximum sensitivity is obtained at the centre zero with a rapid decrease of sensitivity away from the centre zero. This has several advantages; firstly it increases the sensitivity where it is most needed, secondly it protects the moving coil system from large forces in the event of large out of balance currents flowing and thirdly the galvanometer does not go so easily right off the scale but indicates that balance is being approached even when considerably off balance. The required properties for the galvanometer are - a coil resistance of about 100 ohms and a minimum sensitivity near the centre of about 0.1 mm. per microamp. The damping should be approximately critical near the centre zero.

d) The resistances R_1, R_2, and R_3 of Fig.I are of constantan double silk covered, and are wound on flat pieces of paxolin shown at 7 of Fig.2. 4 volts are normally used across the bridge, and with this voltage about 0.2 watts is being continuously dissipated in both R_2 and R_3. These resistances are therefore wound with open spacing to reduce the temperature rise. The resistance Z of Fig.I is purely a ballast resistance to permit driving the bridge from the aircraft accumulator. It is about 60 ohms for a 12 v. supply or 160 ohms for a 24 v. supply. Its thermal dissipation is quite large, so it is mounted externally. With a 12 v. supply we have used a 12 v. 2.2 watt bulb, its light acting as a warning that the instrument is switched on. It may also be mentioned that with 4 v. across the bridge there is a dissipation of about 0.02 watts in the resistance R_1 of Fig.I and also in the thermometer element. This dissipation is not sufficient to affect the element; at about 100 m.p.h. it causes a temperature rise of about 1/4°F.

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7. The Slide Wire

a) One difficulty that was encountered in the development of the instrument was the material of the slide wire. It has a maximum length of about 9" and if its resistance is low the total resistance through R_2 and R_3 must be correspondingly low, and the current through the bridge is undesirably high. On the other hand the wire is subject to wear due to the sliding contact and to avoid trouble from this it must be hard and of the maximum possible cross section. Constantan is fairly satisfactory, but it is relatively soft, and oxidises slowly so that perfect contact is not always obtained. It may be noted that imperfect contact causes troublesome flickering of the galvanometer but does not otherwise affect the result. The most satisfactory material yet used is Staybrite. In the hard drawn condition it is very hard, or if the annealed wire is used it work hardens so rapidly that the sliding contact will harden the surface on which it acts. It has a high specific resistance, twice as high as constantan, and remains bright even under adverse conditions. In the freshly drawn condition it has a thin oxide coat on it, but this is very rapidly removed with a few strokes of cotton wool wetted with metal polish, and good contact is then continuously obtained. Unfortunately Staybrite possesses a temperature coefficient just sufficient to introduce perceptible error if the bridge box temperature changes by 25°C. from that during calibration. Because of this it is possible that Nichrome (80% Ni. 20% Cr.) may ultimately prove a more suitable material for the wire. The sliding contact is a flat piece of phosphor bronze with a short length of constantan wire soldered to the under surface so as to make a rounded edge contact. Silver-gold contact alloy should, however, be more satisfactory than constantan and is recommended for general use.

8. The Scale.

a) The scale is engraved on a 4" diameter dial covering about 9/10ths of the total circumference, and reading is against a single fiduciary mark. The scale is directly coupled to the arm which carries the sliding contact. The dial is clearly seen at 8 in Fig.2, and the stud on which the fiduciary mark is engraved is shown at 9. Externally the dial dominates the instrument, as can be seen from the photograph of the prototype Fig.3

b) The scale is not linear; 1°F. divisions are about 1.3 mm. apart at the high temperatures and 2.1 mm. apart at low temperatures, so that the necessary openness of scale referred to in Section 2b has been achieved. Though a linear scale has advantages, there is also considerable merit in having the instrument with a more open scale and more sensitive at the low temperatures, as observations made at low temperatures will be made by a pilot suffering more or less from the effects of great altitude. Ideally the sensitivity of the galvanometer should be such that near balance the movement of the pointer per °F. change of temperature is slightly more than the length of scale corresponding to 1°F. so that the accuracy of balancing the bridge can be slightly more than the reading of the dial. The sensitivity of the instrument used here falls slightly short of that ideal, and it is the accuracy with which the bridge can be balanced which limits the sensitivity of this apparatus. In the laboratory, if an external very sensitive galvanometer is used the rest of the apparatus permits repeatable adjustment to the accuracy of reading of the bridge, which is about 1/10°F. and the apparatus is unaffected by gentle hammering with a light mallet.

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9. The effect of loss of direct reading

It was with some hesitation that a non-direct-reading instrument was developed for use in single seater aircraft so that some comment is required as to its effect. The bridge is controlled through a 9-1 ratio indirect drive, the controlling knob being a large ebonite knob seen concentric with the scale in the photograph Fig. III and gripping it through thick gloves is not difficult. The indirect drive makes delicate handling of the knob unnecessary and the dead beat galvanometer behaves as if chained to the knob so that balancing the bridge to the nearest whole °F. is an easy operation. The instrument has been installed in a Spitfire and used in regular ascents in H.A.F, The result of this experience is that even the most critical pilots prefer to balance the bridge and obtain the advantages of the long scale to the difficulty of attempting an accurate reading on the short scale of a direct reading instrument. Another advantage lies in the fact that the bridge may be balanced and then the result recorded at leisure while climbing to the next level.

10. Speed of Response

a) The lag of the instrument is partly due to lag in the galvanometer and partly due to lag of the sensitive element. The lag coefficient of the bulb is very small, one or two seconds at speeds greater than 100 m.p.h. as measured in a wind tunnel. The galvanometer is critically damped and has a period of about ½ sec. so that if the knob can be rotated with sufficient speed a reading can be made in less than 2 seconds. The lag of the complete instrument is too small to measure in the air. The test applied consisted in flying level at a low speed then diving through about 2,000 ft. and levelling off again into steady flight at the high speed at which the aircraft is flying after the dive. The thermometer is read as soon as possible after level flight has been attained but it is already steady. Thus by the time the observer has recovered from the accelerations of the dive, the thermometer is in equilibrium to the accuracy of measurement (about 0.2°F). Since the thermometer can be subject to a temperature change of about 20°F. in 20 seconds in this way, the speed of response must be very rapid.

b) This rapidity combined with the openscale is very useful during ascent to indicate the base or top of an inversion. It is at once obvious when an inversion or isothermal layer is encountered.

11. Flight Tests

a) Flight tests have been made to assess the actual accuracy obtained by the thermometer, its suitability for routine use, and its reliability under aircraft conditions of service. The instrument in the version described here, and also in earlier versions, has been used in many flights. Results employing mounting of the element without a radiation shield were given in a previous report Met. (H.A.F.) R/4 and the indicator was used in a Spitfire aircraft to obtain the results quoted in the report Met. (H.A.F.) R/7 giving the speed correction to a thermometer mounting which was described in that report. The results were fitted to the formula $T = T_o + \alpha V^2$ by the method of least squares where T is the indicated temperature, T_o is the temperature of the air, α is the speed correction coefficient, and V the true air speed. The results are repeated in Table I. Similar results have been obtained for Mk.II and Mk.V. psychrometer and the results computed by Professor H. H. Plaskett, F.R.S. It is instructive to compare the results of the six psychrometer runs with the results obtained by the balanced bridge and these are also quoted in Table I. The mean probable errors for the strut psychrometers and the balanced bridge are not very different. The probable error for the strut psychrometer is slightly greater than that of a single reading, between 0.1 and 0.2°F., Since this is the error of the average of about 7 readings the cause of the error must be lack of homogeneity of the atmosphere, and we conclude that the accuracy attained by a Spitfire pilot using a balanced bridge is about as high as atmospheric homogeneity justifies. Actually the bridge used was an early, not very well made, model and its reproducability of its readings was not as good as the bridge described here, though, as the figures show, it was good enough.

Table I.

Date	Aircraft	Press.mb.	T. °F.	Prob. error in T. °F.	α	Prob. error in α
		Balanced Bridge.				
6.2.43	Spitfire P.7301	960	41.5	0.21	1.42	.10
8.2.43	" "	875	26.6	0.21	1.59	.09
12.3.43	Spitfire AB.450.	955	37.4	0.13	1.43	.05
14.3.43	" "	590	0.9	0.26	1.55	.08
"	" "	934	37.2	0.15	1.69	.06
15.3.43	" "	369	-43.3	0.36	1.53	.12
	Mean probable error			0.25		.095
		Strut Psychrometer Mk.II.				
21.1.43	Boston AL.480	920	42.7	0.10	1.545	.043
14.2.43	" AL.481	973	48.9	0.27	1.303	.114
27.2.43	" AL.481	960	37.7	0.20	1.522	.081
		Strut Psychrometer Mk.V				
21.1.43	" AL.480	920	42.3	0.12	1.117	.051
14.2.43	" AL.481	973	48.0	0.20	1.035	.082
27.2.43	" AL.481	960	36.9	0.14	1.120	.060
	Mean probable error.			0.20		.082

b) The results quoted here and discussed in 11a concern only casual errors including errors of reading, and do not show any systematic errors. Provided the speed correction is known accurately systematic errors should not arise. In order to obtain the data to permit engraving the scale, the instrument was calibrated, using a dial engraved in angular degrees, against a thermometer taken from a strut psychrometer which had an N.P.L. certificate giving its corrections to 0.2°F. The results were then used to calculate the angular position of each mark on a Fahrenheit scale, and from these data a Fahrenheit scale was engraved on a blank dial. This scale was assumed correct, though slight errors in the engraving were known to require a correction of about -0.3°F. at the upper end of the temperature scale, and this was confirmed by a quick room temperature check.

c) The thermometer graduated in °F. was then mounted in a Spitfire (AB 450) and used on temperature climbs for about three weeks. The aircraft became unserviceable and for about six weeks the thermometer remained installed while the aircraft was extensively overhauled, and it was finally used on another ascent. It was then transferred to a Boston aircraft and carried on several further flights. To test whether the thermometer had been adversely affected by this treatment it was recalibrated with the following results :-

Table II

Indicated Temp.	Correction to be applied
-69.1°F	-0.1°F.
-63.9	-0.1
-53.7	-0.1
-36.0	-0.2
-27.2	-0.2
-20.6	-0.2
-10.1	0.0
- 0.3	+0.1
8.3	+0.[illegible]
19.9	0.0
30.6	-0.3
49.8	-0.3
48.7	-0.2
61.7	-0.3

The accuracy of the test is limited by the sensitivity of the galvanometer to about 1/4°F. and since a negative correction was expected at the upper end of the scale the results are highly satisfactory.

d) On account of the temperature coefficient of the staybrite wire used for the slide wire, the indicator will possess a small temperature coefficient. This temperature coefficient is only just large enough to be significant, as will be seen from the table below which gives the approximate calculated correction which would be required if the temperature of the indicator were 50°F. below the calibration temperature.

Table III

Indicated Temp.	Correction to be applied
+70°F.	+0.8°F.
+30	+0.4
0	+0.1
-40	-0.3
-80	-0.7

e) Observations have been made of the temperature of a normal Spitfire cockpit during vertical soundings to about 40,000 ft. The lowest temperature recorded was 30°F. and it is unlikely that the inside of the indicator would fall as low as that. Thus normally the temperature coefficient correction would only be about half the corrections quoted above. Nevertheless it may be possible to improve on this by using Nichrome (80%. Ni.20%. Cr.) for the slide wire.

12. Conclusions.

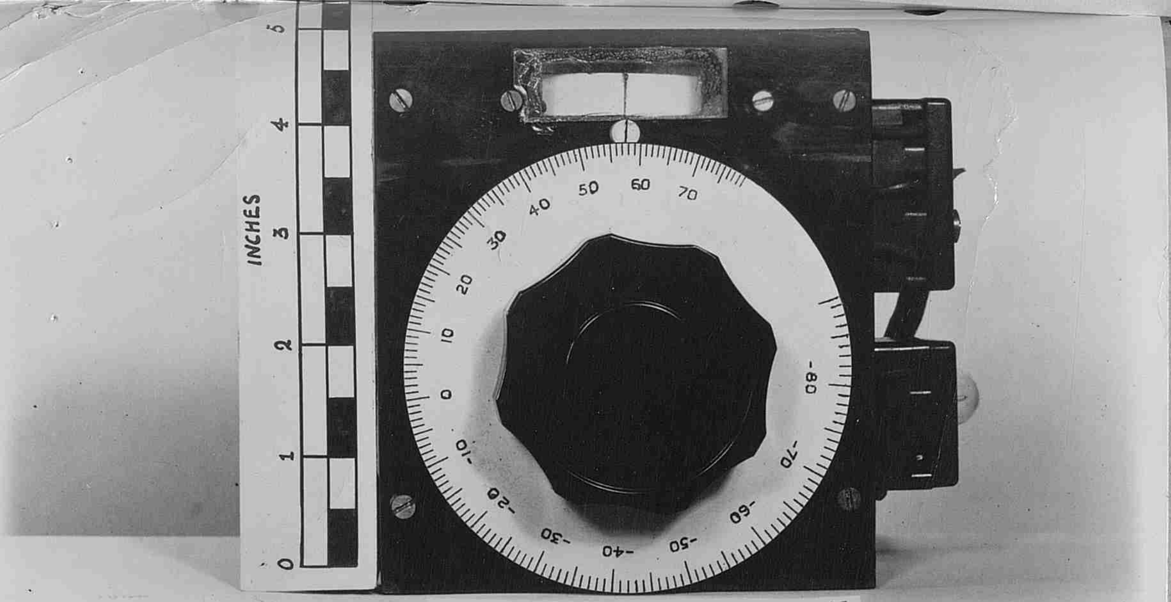
It appears that the instrument is capable of satisfying the requirements laid down in Para.2.

13. Further Work

a) The galvanometer is at present, as regards its sensitivity, the weakest point in the instrument and the question of obtaining a better galvanometer is being followed up. By slight redistribution of parts, it is [illegible] shock mounting for the galvanometer without increase in the dimensions and this will reduce the difficulties of manufacture of the galvanometer.

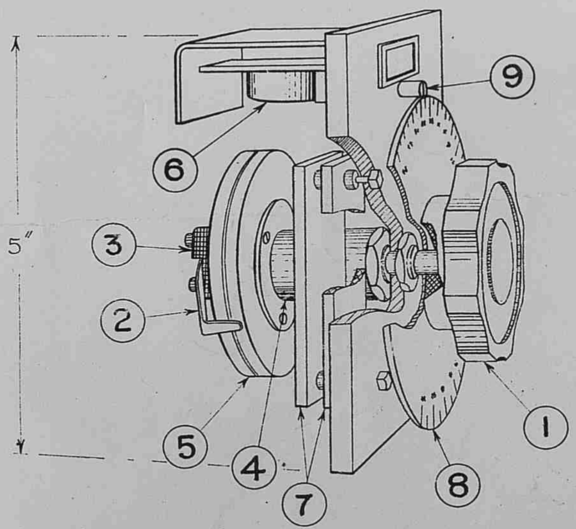
b) Slight improvement is also possible by reducing the temperature coefficient of the slide wire. As has been mentioned Nichrome (80% Ni. 20%Cr.) offers most hope in this respect and it is hoped to make a trial in the near future.

c) The element used here has been the Cambridge Instrument element. The possibility of using a standard A.M. temperature element which is more widely available is being investigated.



SCALE MARKED in DEGREES F.

FIG.2



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i) The Cambridge aircraft thermometer consists of a platinum resistance thermometer in a bridge circuit which is balanced for a temperature of (usually) -20°C. The actual temperature is indicated by a moving coil 2-pivot indicator with centre zero operated by the current which flows in the 'galvanometer' link of the Wheatstone's bridge, due to the bridge not being balanced.

The element consists of a fine silk covered platinum wire doubled back upon itself and then drawn through a thin copper tube. The copper tube is wound in the form of a helix on a suitable former with spacing wires between turns of the helix. The element is mounted on the under surface of the wing of the aircraft with the axis of the helix pointing forward. Connection to the bridge is by means of uncompensated leads. The bridge is built into a control box which is normally mounted somewhere in the cockpit of the aircraft. Provision is made for adjustment to take up variations of battery voltage, the indicator when off balance being dependent upon battery voltage.

THERMOMETER - NOTES ON ITS USE

by A.W.BREWER. M.Sc.

Summary.

The position regarding the accuracy of the air temperatures as measured by Meteorological Spitfire aircraft is briefly reviewed. The main sources of error, other than actual change of calibration or mechanical breakdown, are discussed and suggestions made for eliminating some and reducing the effect of others.

1. Accuracy of the Temperatures as measured on High Level Spitfire Ascents.

The only yardstick available for checking the accuracy of the temperatures as measured by the Cambridge Electrical Resistance thermometer commonly used on high level Spitfire (Prata) ascents, is the temperature as measured by simultaneous Gladiator ascents using a strut psychrometer. Statistical comparisons have been made by M.O.2.B.S. and by Dr. Dobson which show that in addition to a random difference which averages about 3 or 4°F. there is a consistent difference, the Prata temperature being on the average 2.5 to 4°F. lower according to the level of the observation. Also a comparison made at Wick between a Strut psychrometer and a Cambridge Electrical Thermometer both mounted on a Hudson aircraft showed that quite large differences could appear; the largest difference observed was 8°F. These observations were discussed at a Meteorological Office conference on upper air temperature measurement held at Adastral House on 7th October, 1942, and the evidence was presented to that Conference in detail. As the required standard of accuracy is 1°F. these errors are serious and the object of this note is to discuss their cause and eliminate or reduce them as far as practicable.

2. The Sources of Error.

The main sources of error to an aircraft thermometer are:-

i) Errors in the indicator due to vibration etc.
ii) Errors of reading, including parallax.
iii) Difficulties with the element including the effects of insolation and rain.
(iv) Application of the wrong speed correction.

These sources of error are dealt with in turn below.

3. Errors in the Instrument.

i) The Cambridge aircraft thermometer consists of a platinum resistance thermometer in a bridge circuit which is balanced for a temperature of (usually) -20°C. The actual temperature is indicated by a moving coil 2-pivot indicator with centre zero operated by the current which flows in the 'galvanometer' link of the Wheatstone's bridge, due to the bridge not being balanced.

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The moving coil indicator is mounted on the instrument panel of the aircraft. It is of the edge-wise pattern and the scale is curved through about 90°. The pointer is a knife edge with a mirror behind, which on the ground permits the effect of parallax to be eliminated. The scale is 44 mm. long and normally covers the range -20°C. to -70°C. On the ground the scale can be read with confidence to about $\frac{1}{2}^{\circ}$C. but in the air one usually find difficulty in reading the scale to better than 1°C.

ii) Effect of Vibration.

Apart from general deterioration this also affects the indicator. Waves are set up in the hairsprings which control the moving coil and can either result in a steady parasitic force which makes the reading incorrect, or it can cause the pointer to swing, usually over a range of 5 or 10°C. It can also set up resonance in the pointer, making it vibrate and appear as a broad pointer instead of a sharp knife edge. The first effect is difficult to detect. The other effects are immediately obvious and can usually be stopped by changing engine speed. If for any reason the first effect is suspected, it may be detected by changing engine speed. The indicated temperature should not change provided speed and height remain constan Whenever possible the engine speed should be reduced by coarsening the propeller pitch.

iii) Effect of Battery voltage variations.

Since the reading depends on the out of balance current, the reading will depend on battery voltage, particularly at end of the scale but provided batteries in good condition are used, neigher too freshly off charge nor too near exhaustion, little trouble will be experienced in this respect.

4. Effect of parallax.

As has been remarked above, the scale is considerably curved, which greatly increases the possibility of parallax error though a mirror is provided to facilitate its elimination. Unfortunately in the restricted space of a Spitfire cockpit it is not possible to place the eye in the required position when the pointer is at the end of the scale and the possibility of error arises. There, should, however, be no real trouble in this respect, for if the eye cannot be placed in the correct place a reasonably accurate reading can be obtained by taking the point in the scale half way between the apparent position of the needle and its image in the mirror.

5. Difficulties with the element and its mounting.

The element, unless grossly mishandled, should be satisfactory, but the mounting as used at present on Prata aircraft is subject to several objections.

The electrical connections are taken by means of soldered contacts to connections at the back of the element. In rain, or drizzle, water which deposits on certain parts of the element will be swept back on to these connections and cause electrolytic effects which may give spurious temperature readings. This can be eliminated entirely by the use of a small waterproof cover which has been designed by Mr.Hartley of M.O.4.

Also, though the element is shielded from downward radiation by the aircraft wing, these measurements are often made above cloud, where the upward flux of radiation must be considerable. Trials made at Boscombe Down showed that at 200 m.p.h. insolation at 20,000 ft. raised the temperature of a thermometer by $1\frac{1}{2}^{\circ}$C. The reflexion coefficient of clouds may rise to 80% which would cause a thermometer t read about 1°C. high. The inclusion of a radiation shield is therefore advisable.

These two features have been incorporated in a new mounting.

The element with waterproof box is mounted inside an aluminium radiation shield, the whole thing being built on to the same inspection hole cover as is normally used to carry the element on Prata aircraft.

6. Speed Correction Coefficient.

At present a speed correction coefficient of 100% i.e. ($T = 1.0 \left(\frac{V}{100}\text{ mph}\right)^{2\circ}$C. or $1.8 \left(\frac{V}{100}\text{ mph}\right)^{2\circ}$F.) is being applied. The systematic difference between Prata and Thum temperatures points to an error in the speed corrections, since this difference increases as the speed correction increases, and is consistently about 30% of the speed correction.
Preliminary observations on the Boston Aircraft at Boscombe Down confirmed that the coefficient was probably appreciably less than 100%. Further tests were made when a balanced bridge was available permitting more accurate measurement of the temperature of the Cambridge element. Results obtained showed that the coefficient varied somewhat with the mounting according to interference by other bodies, and that in a free airstream the coefficient was about 65%

A replica of the Prata mounting was made on a Boston aircraft and its correction determined by measuring its temperature on a balanced bridge and comparing the reading with that of a Mk.II. psychrometer mounted on the aircraft.

Excellent results were obtained giving a speed correction for the Prata replica of 13% less than the Mk.II. psychrometer, i.e. 67% Three runs were consistent as regards slope but showed a general tendency for the Prata element to read high by amounts up to 1.4°F. The effect was attributed to radiation since a run made below cloud near sunset showed no such effect.

The element was therefore incorporated in the new mounting described in para.5. above as this is most likely to be free from criticism, and mounted on a Spitfire. The balanced bridge, which had behaved well, was also rebuilt so as to be suitable for use by the pilot in a single seater.

Speed Trials.

Speed trials have been carried out by F/Lt. G.H.Proctor with this equipment. The method is the usual one of flying at a selected level, and reading the thermometer temperature at various speeds between minimum and maximum speeds. The observations have all been made under unstable atmospheric conditions. This has two advantages: firstly the mixing which takes place ensures a higher degree of homogeneity in the atmosphere, and also the effect of position errors on the Altimeter and air speed indicator oppose each other and reduce the effect to such an extent as to render the application of any corrections unnecessary.

The main difficulty in the assessment of the results is the lack of homogeneity in the atmosphere; to eliminate this, six trials at levels between 1,500 and 25,000 ft. have been made. The results have been fitted to the form $T = T_0 + \alpha \left(\frac{V}{100}\text{ mph}\right)^2$ by the method of least squares, and the probable error calculated to permit assessment of the accuracy of the results and to determine whether differences between observations are significant.

MPR - 113

Results of speed trials.

The results of the speed trials are given in the table below.

Serial No.	Date	A/cs (Spitfires)	Temp.	Press.	α	Prob. error in α
1.	6.2.43	P.7301.	41.5°F.	960 mb.	1.42	.10
2.	8.2.43	"	26.6	875 "	1.59	.09
3.	12.3.43	AB.450	37.5	955 "	1.43	.05
4.	14.3.43	"	0.9	590 "	1.55	.08
5.	"	"	37.2	934 "	1.69	.06
6.	15.3.43	"	−43.3	369 "	1.53	.12

The mean is 1.53 $\left(\frac{V}{100}\text{ mph}\right)^2$F. = 85% of the kinetic energy. The probable error of the mean is .04 $\left(\frac{V}{100}\text{ mph}\right)^2$°F. or about 2%

Fig. 1. and Fig. 2 are graphs of runs No. 1 and 6 in which the observed temperature is plotted against V^2 and they show how the lack of atmospheric homogeneity affects the results.

In Fig. 1. which is a graph of run No.1., the line of slope 1.42 $\left(\frac{V}{100}\right)^2$°F. obtained by the method of least squares is shown as a full line, but it may be noted that all points except one lie within $\frac{1}{4}$°F. of the dotted line which has a slope 1.53 $\left(\frac{V}{100}\right)^2$°F. which is the final mean of all the runs. It seems probable that for the high speed observation the aircraft chanced to find a colder region than for the rest of the observations and caused this point to be low on the graph. This results in a lowering of the slope of the best line and a considerable increase in the probable error of the result.

In Fig.2. which is a graph of run No.6. it is very clear that the points can be divided easily into two groups, as indicated by the two dotted lines. Each has approximately the same slope but the lines are separated by about 1.5°F. suggesting that two regions of air were actually involved in the observations. The distribution of the low points is such as not to affect the resulting slope but it greatly increased the calculated probable error of the slope.

It may be mentioned that this value of the speed correction intermediate between 100% and that of the element in the free air is obtained because the tube of the radiation shield combined with the obstruction of the waterproof box acts as a very leaky forward facing pitot and there is a certain degree of stagnation round the element. This value of the speed correction will only be uniformly attained if the proportions of the system are kept constant to a reasonable degree. The radiation shield should have an area of cross section equivalent to a 3" circle and the diameter of the front flange of the waterproof box should be $1\frac{1}{4}$".

The Balanced bridge which has been used in these trials is suitable for routine use and will be made the subject of a separate report.

Acknowledgments.

I have pleasure in acknowledging help given by Prof. Plaskett, F.R.S. in the technique used in the examination of the observations, and also my debt to S/Ldr. Hastings A.F.C., A.F.M., who has permitted us to carry out this work on H.A.F. Spitfires.

Boscombe Down.

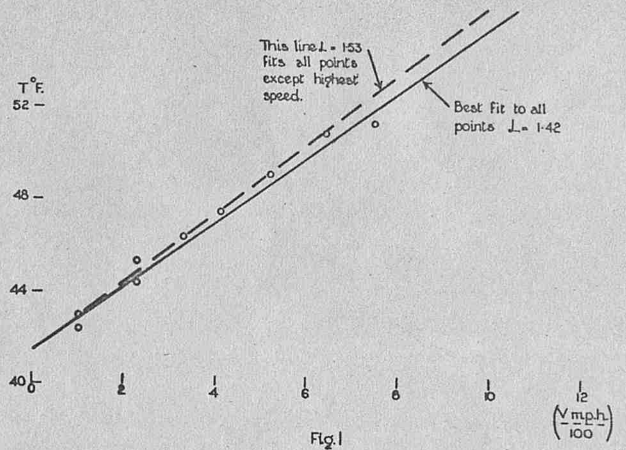
This line L = 1·53 fits all points except highest speed.
Best fit to all points L = 1·42
T°F.
52
48
44
40
0
2
4
6
8
10
12
(V m.p.h. / 100)

Fig.1

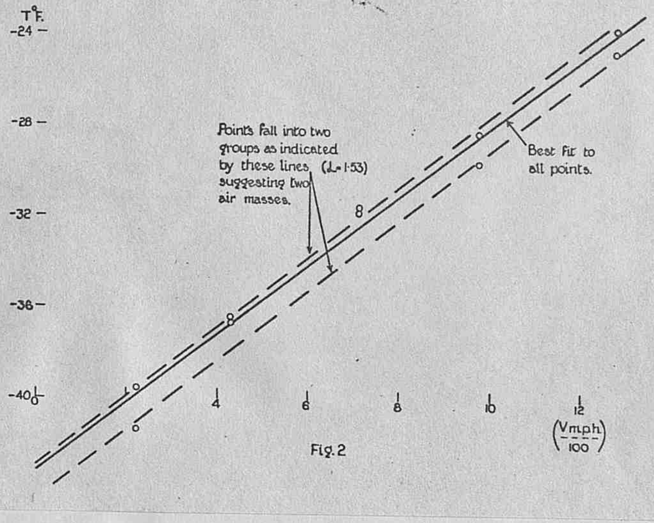
Points fall into two groups as indicated by these lines (L = 1·53) suggesting two air masses.
Best fit to all points.
T°F.
-24
-28
-32
-36
-40
0
4
6
8
10
12
(V m.p.h. / 100)

Fig.2

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MRP115. (revised)
3rd March, 1944.

METEOROLOGICAL RESEARCH COMMITTEE

A New Method of Applying a Photo-electric Cell in the Measurement of Short Period Variations in Wind Velocity.

by A.R.LAIRD, Meteorological Office,
Aberporth.

Note. Comments by Professor G.I.Taylor (MRP 132) and Mr.L.F.G.Simmons (MRP 137) upon MRP 115 have been included in this paper.

1. The Anemometer and Photoelectric Cell.

A 3-cup anemometer has been designed which is very responsive to short period variations in speed in all but light winds, and for use with it a distant recording system has been devised such that an analysis showing these variations may be undertaken. The anemometer is not intended for the measurement of mean winds over long periods.

No gear train or friction producing contact maker is employed in recording the revolutions of the cup head. Instead an arrangement has been made so that the intensity of a beam of light falling on a photo-electric cell varies with the rotation of the cup spindle. The cell controls the flashing of a small neon lamp which is photographed on a continuously moving film. A feature of the light system is that the addition of inertia due to a shutter has been avoided by cutting a slot in the spindle and installing the light source close to the slot so that the spindle acts as a shutter and interrupts the light twice per revolution, or, roughly, once every second for each mile per hour of wind speed.

In constructing the anemometer the results were assumed of the experiments carried out by Sheppard on those cup forms and cup rotor dimensions which give a linear relation between cup speed and wind speed. The same relative dimensions were preserved for the cup head designed for the new anemometer. The internal diameter of the cups is 2.5 cm., with the cup centre 3.5 cm. from the centre spindle. The upper bearing is a miniature ball race spun into the spindle housing. A spring relieved jewel bearing takes the thrust. Steel is used for the bearings at each end of the spindle, duralumin for the spindle itself, and aluminium for the cups.

The slot in the centre spindle is 1 cm. long by 0.15 cm. wide and a 4 v. pea lamp and a small photo-electric cell are mounted on either side of it. The p.e. cell is masked so that only light passing through the slot can fall on the cathode and is directly coupled with a miniature valve as shown in the circuit diagram. Lamp, cell and valve are enclosed in a metal box, 7 cm. high, through which the cup spindle passes. Clearance of twice the height of the case is allowed between it and the cup rotor. The internal wiring terminates on a standard 5-pin valve holder let into the bottom of the case.

If a further reduction in size is required, a smaller p.e. cell will reduce the case dimensions and make possible the use of a shorter cup spindle, but the weight of the moving parts (which, in the model described above, is just over 7 gm.) can most usefully be reduced by decreasing the thickness of the aluminium used for the cups.

....

2. The Recording Arrangements.

The method used to control the flashing of the neon lamp is very simple and is shown in the circuit diagram - Fig.1. The neon lamp is connected in parallel with the valve, with the anodes of each connected to opposite plates of a condenser and then each through a large resistance to the same H.T. supply. The voltage is adjusted to be sufficient to cause the neon lamp to strike so that normally it is glowing. The grid voltage of the valve is adjusted to be less than the cathode voltage when no light (other than a very little scattered light) is affecting the p.e. cell. When light falls on the p.e. cell the grid voltage of the valve changes, the valve passes current, and the potential drop across it is decreased. The potential difference across the neon lamp varies in sympathy so that it falls below the striking voltage for the lamp, and the lamp is extinguished until the condenser becomes charged or until current stops flowing through the valve. With the values of the components shown, the valve controls the neon lamp completely even in anemometer speeds as low as one half revolution per second. Distant recording over a distance of 400 yards has been successfully undertaken using the above arrangement.

The trace obtained consists of a series of dashes on the film the corresponding ends of successive dashes marking a half revolution of the cup spindle. A second neon lamp flashing with A.C. mains frequency is mounted alongside the anemometer neon lamp and is photographed with it to provide a time base independent of the smoothness of the film speed. Recording seems to be trouble free, but as all the electrical components are new this is only to be expected. The small number of components and the simplicity of the circuit should make it easy to locate any faults that may arise.

The method of recording has the disadvantage that measurement of the trace for equal intervals of time is inconvenient. In practice the time interval for an integral number of extinctions of the neon lamp is measured, the number varying so that the time lies between 0.15 and 0.2 seconds. The probable percentage error is wind speed introduced in measurement of the trace will therefore be roughly constant and less than 3% of the measured wind speed.

3. Performance.

Tests of the anemometer in a wind tunnel in a mean wind of 15 f.p.s. with a regular variation superimposed, showed that for a fluctuation period of 1.5 sec. the recorded amplitude was 60% of the true amplitude. On open sites the ratio of amplitude to mean wind speed in eddies of this period is not likely to be large so that the error in the wind recorded will be small. Test runs to indicate the scope of the anemometer in natural winds have been made with the instrument installed near the edge of the flat roof of a building so that gustiness was very great. The records show that the anemometer can resolve easily gust maxima with a separation of 1-1/4 seconds or more, and can indicate the passage of an eddy of 3/4 second period.(See Fig.2).

The following theoretical discussion of the performance of this instrument has kindly been supplied by Professor G.I.Taylor:-

"Regarded as an instrument for measuring gustiness the instrument has one interesting characteristic in which it differs from all other turbulence measurers. It has no period or time of lag which is characteristic of its mechanism. It has a lag which is probably very large for a gustiness or turbulence recorder, and is responsible for the large error of 40% and a $1\frac{1}{2}$ second pulsation; but this lag is inversely proportional to the mean wind speed. This means that if you set the instrument up in the eddies produced by say a roof or wall the amount by which it underestimates the percentage gustiness will be the same at all wind speeds, assuming that is, that the linear scale of the turbulence pattern is the same at all wind speeds. One can write down a simple equation to represent its action in a varying wind. If U is the mean wind velocity and u the variation from that mean, Ω the mean angular velocity of the cups and ω the variation in angular velocity, the equation of motion of the cups may be written:-

$$I\frac{dw}{dt} = K(u - m\omega)U \qquad 1.$$

where $m = U/\Omega$ is the constant relating the velocity of rotation to the wind speed when steady, and K is a constant depending on the aero-dynamic properties of the cup sustem. I is the moment of inertia of the cups.

If you now consider the behaviour of this anemometer when placed in a wind with a given linear scale of turbulence but varying values of U you can represent the velocity variations by the equation:-

$$u/U = f(Ut) = f(x) \qquad 2.$$

where x is written for Ut. Putting this value of u in (1) one gets:-

$$I\frac{d}{dx}\left(\frac{\omega}{U}\right) = K\left(f(x) - m\frac{\omega}{U}\right) \qquad 3.$$

so that ω/U is a function of x only and is independent of U. The instrument therefore records fluctuations which are the same percentage of the true fluctuations for all mean wind speeds.

If a simple harmonic fluctuation occurs one can put:-

$$f(x) = a \sin pt$$

and the well known solution of (3) shows that the ratio

$$R = \frac{\text{recorded fluctuation of wind}}{\text{true fluctuation of wind}} = \frac{1}{\sqrt{1 + \frac{p^2I^2}{m^2K^2}}}$$

If one takes the measured value of R = 0.6 when U = 15ft/sec. and period =1.5 secs. so that $p = \frac{2\pi}{15 \times 1.5}$ one gets

$$\frac{p^2I^2}{m^2K^2} = \frac{1}{(0.6)^2} - 1 = 1.78$$

This enables you to find out what the instrument would record if the period were shorter. Thus if the period of fluctuation were only 0.5 secs instead of 1.5 secs in a 15ft/sec wind the instrument would only record 25% of its true magnitude."

Professor Taylor further points out that the instrument can have only a limited value for measuring the correlation between velocities at neighbouring points because the actual values of the turbulence components are too much in error (40% error in a fluctuation of period 1.5 second and 75% error in a fluctuation of period 0.5 second).

With regard to the practical drawback that some time must elapse after taking a run of the wind before a useful record is obtained, Mr. L.F.G. Simmons has suggested ways of overcoming this. He proposes the addition of either an electronic frequency meter, or a counting device similar to that employed for measuring the rate of small electronic pulses produced by gamma radiation.(1)

By such means the rotational speed of the anemometer would be registered by the output current and thus any ordinate of a continuous record of the current would give a measure of wind speed. Whether such an extension of the present circuit would be justified depends upon the purpose which the instrument is intended to serve.

The instrument may be found useful for obtaining short term records of variations in natural winds especially in positions where photography of the pointer and scale of a direct reading anemometer is difficult or where installation of a hot wire or other more sensitive anemometer is inconvenient, i.e. where there is a lack of facilities for check calibrations. Traces from a number of such anemometers can be obtained simultaneously on the same strip of film so that if the attenuation of the turbulence pattern is acceptable, correlation of wind speed at different points can be attempted.

4. Acknowledgements. In addition to the suggestions received from Professor Taylor and Mr. Simmons mentioned above, the author wishes to acknowledge the help and advice given by the Meteorological Section, Porton, and by the Electronics Section, Aberporth.

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(1) A direct-reading counting rate meter for random pulses. N.S. Gingrich, R. Evans and H.E. Egerton, Rev.Sci.Inst., 7, p.p. 450-456, 1936.

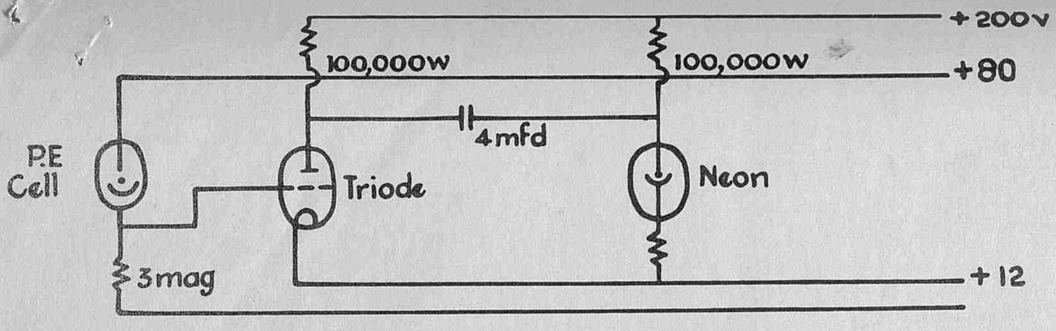

+200v
100,000w
100,000w
+80
4mfd
P.E
Cell
Triode
Neon
3mag
+12

Fig.I. Circuit Diagram

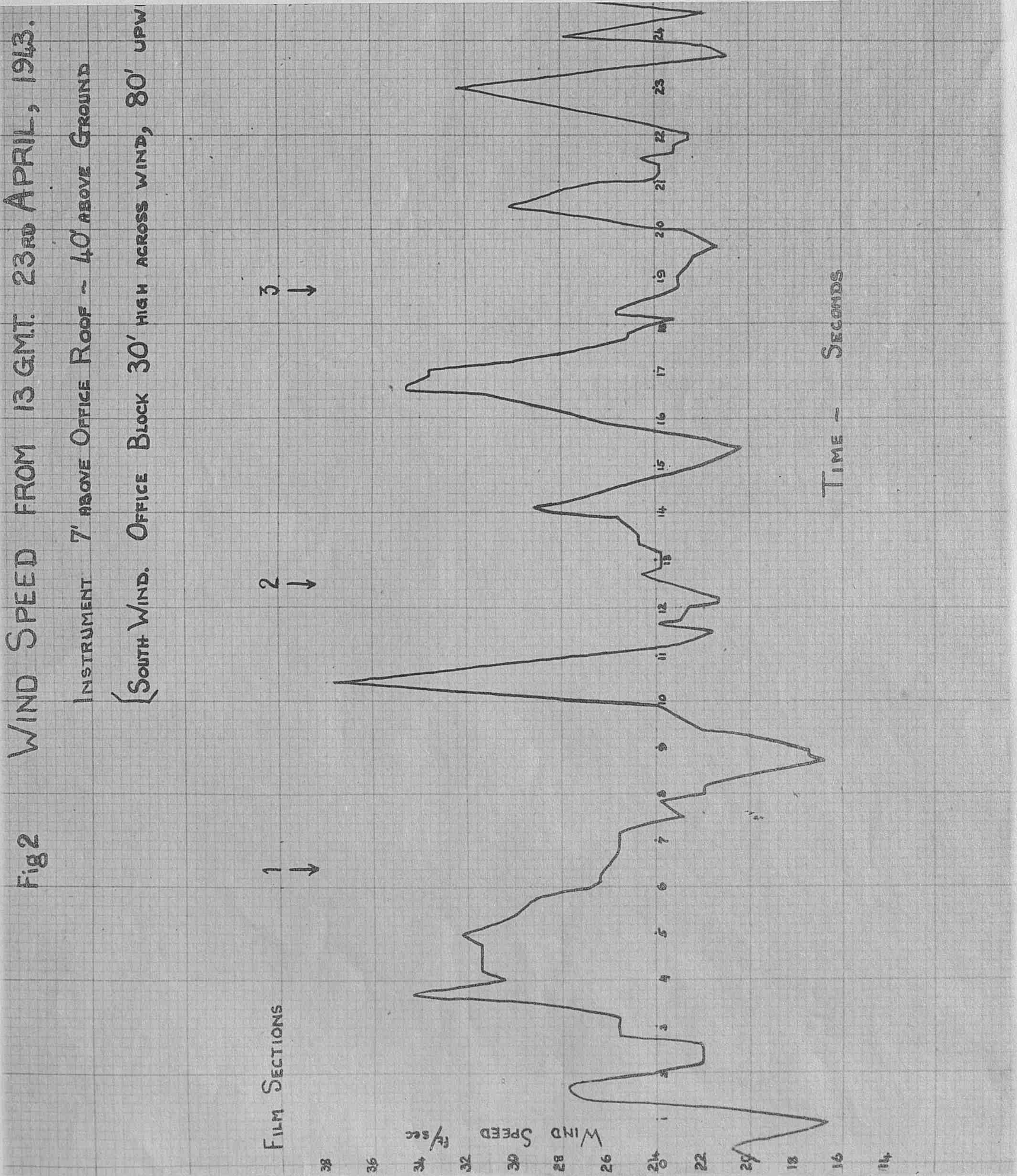


Fig 2 WIND SPEED FROM 13 G.M.T. 23RD APRIL, 1943.

The requirements for these three conditions are taken to be as follows:-

(i) An accuracy of $\pm\frac{1}{10}$°C - convenience if possible.

/(ii)

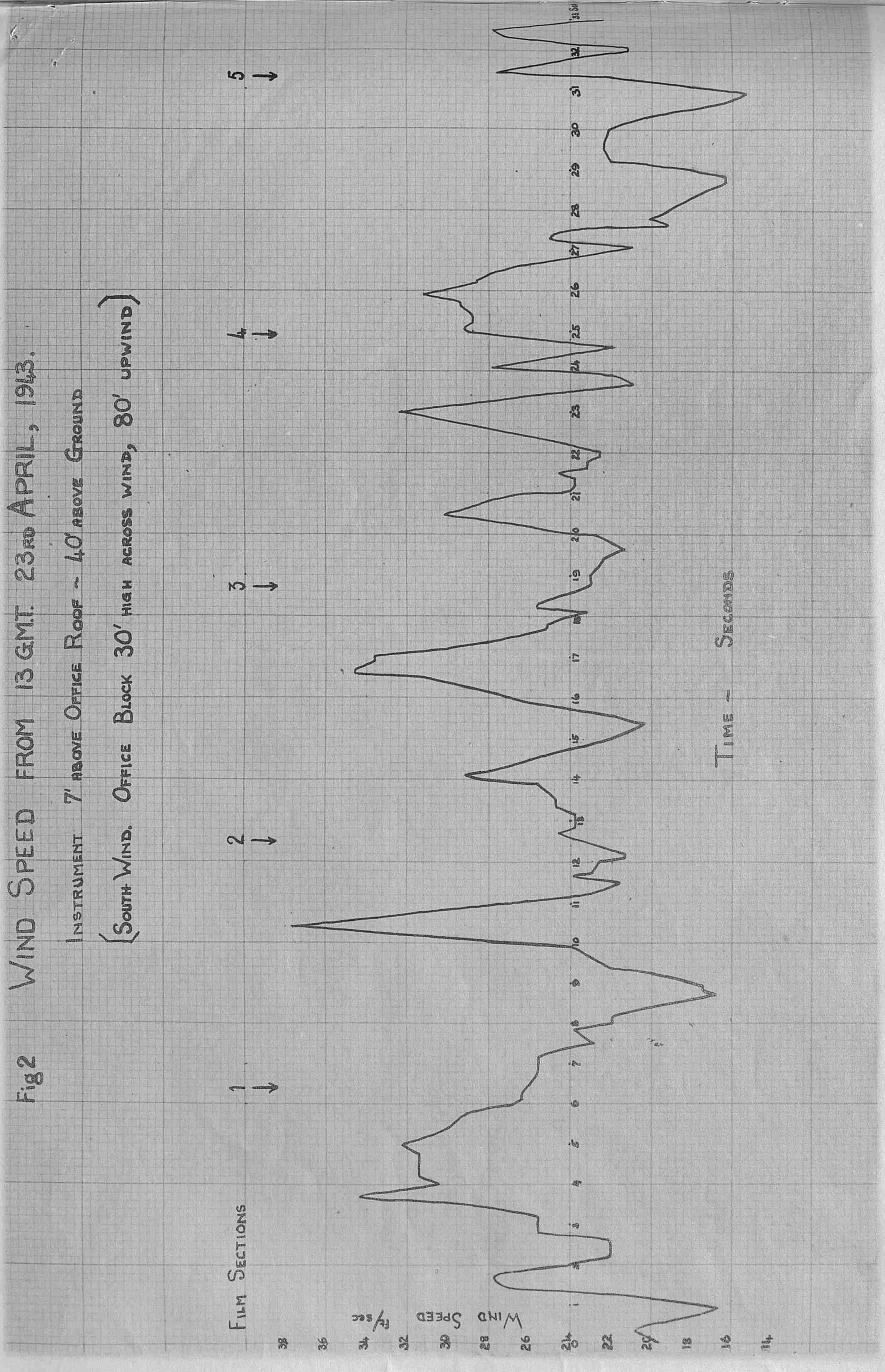


Fig 2 WIND SPEED FROM 13 G.M.T. 23RD APRIL, 1943.

INSTRUMENT 7' ABOVE OFFICE ROOF ~ 40' ABOVE GROUND

(SOUTH WIND. OFFICE BLOCK 30' HIGH ACROSS WIND, 80' UPWIND)

The requirements for these three conditions are taken to be as follows:-

(i) An accuracy of $\pm\frac{1}{10}$°C - convenience if possible.

/(ii)

AIRCRAFT AND ARMAMENT EXPERIMENTAL ESTABLISHMENT

16 MAY 1943

BOSCOMBE DOWN

MRP 117

Brief comparison of air temperature thermometers used and tested at A.& A.E.E. for meteorological work

By
R.M.Goody B.A.

A.& A.E.E. Ref:- 3708/5-CTO/G.32A.
M.A.P. Ref:- Nil.

R & D.

Summary

A number of air thermometers have been tested on various aircraft at this Establishment, and the report deals with these instruments and their suitabilities for the following purposes:-

(i) As research standard instruments.
(ii) As instruments for routine meteorological ascents.
(iii) As standard fittings to operational aircraft.

The following thermometers are considered in this report:-

(A) The spirit thermometer used in a meteorological Strut Psychrometer.
(B) The Cambridge electrical resistance thermometer in three modifications.
(C) The Negretti and Zambra nitrogen filled capillary thermometer.
(D) A balanced bridge electrical thermometer.
(E) An R.A.E. pattern L-shaped spirit in glass thermometer.
(F) An N.P.L. multi-junction thermocouple thermometer.

It is shown that Condition (i) is fulfilled by thermometers (A) and (F), Condition (ii) by thermometers (B) and (D), Condition (iii) partly by thermometer (E), and that thermometer (C) has a useful function as an approximate check in routine meteorological ascents.

1. Introduction.

The following thermometers will be briefly considered in this report:-

(A) The spirit thermometer used in a Meteorological Strut Psychrometer.
(B) The Cambridge electrical resistance thermometer in three modifications, (e.g. with curved indicator and with single and double reading flat indicators).
(C) The Negrette and Zambra nitrogen filled capillary thermometer.
(D) A balanced bridge electrical resistance thermometer.
(E) An R.A.E. pattern spirit in glass thermometer which may be read inside an aircraft (fully reported in A.& A.E.E/Inst/37).
(F) An N.P.L. multi-junction thermocouple thermometer (fully reported in A.& A.E.E./Inst/40).

(A) (B), (C) and (E) were all tested together upon a Boston III aircraft in order to assess their comparative properties. (D) was tested on the same aircraft immediately subsequently to these comparative tests. All the thermometers were mounted around the nose compartment of the aircraft and were read by an observer inside. (B) (C) and (D) are used regularly on meteorological Spitfires whilst (A) is a standard instrument used by the meteorological Office.

The object of this report is to provide a short summary of the experience gained upon all these thermometers, and to consider the suitability of each type for the following uses:-

(i) As research standard instruments.
(ii) As thermometers for routine meteorological ascents.
(iii) As standard fittings to operational aircraft.

The requirements for these three conditions are taken to be as follows:-

(i) An accuracy of $\pm\frac{1}{10}{}^{\circ}$C - convenience if possible.

(ii) An accuracy of ±¼°C - should be easy to read - should be completely reliable with frequent inspection.

(iii) An accuracy of ±2°C - very simple and robust - should be completely reliable and permanent and need no attention.

1. The Meteorological Office Strut Psychrometer.

1.1. Description. There exist several Marks of psychrometer, but in general the thermometers used are all spirit in glass thermometers with a scale about 1 foot long covering an approximate range of +25 to -45°C (different marks having different ranges). The divisions are always 1°F and are marked both on the glass and on the support for the stem. The bulb is housed in some kind of enclosure which both serves as a radiation shield and a ventilating system. The psychrometers upon the Boston were unnecessarily complicated as far as this work was concerned by the fact that the bulb housing was designed mainly for wet bulb ventilation. The psychrometer has to be mounted vertically upon the outside of the aircraft in such a way as to be clear of any exhaust heat, etc. and so as to be read easily from inside. If read through a window this must remain clear.

A photograph of the Mark V psychrometer is appended.

1.2. General properties. The thermometers used in psychrometers are standard thermometers of considerable accuracy, it being possible for a competent observer to read to 1/5th of a division (1/10 °C), and they can be easily calibrated accurately under laboratory conditions. The speed correction varies with the nature of the bulb housing; for the common Mark II psychrometer (honeycomb housing) the correction is 80% of the full adiabatic whilst for the Mark V psychrometer (simplified housing) it is about 55%. The time lag of the thermometer is very large compared with other standard types, 2 or 3 minutes level flight being necessary to ensure complete stabilisation. The size of the psychrometer was sufficient to cause a serious disturbance of airflow on the Boston, resulting in marked buffeting of the aircraft at high speeds. The thermometer is unaffected by moderate vibration and the fitting is very robust.

The above properties make the instrument suitable for use as an accurate standard within the range covered provided a suitable position can be found.

1.3. Further developments. A thermometer of this type with scale extended to -60°C or -65°C would be of great use as an accurate standard.

2. Cambridge electrical resistance thermometers.

2.1. Description. This thermometer consists of three parts, the resistance element, the control box, and the indicator. The resistance element is a thin platinum wire contained inside a copper spiral, it is robust and permanent. The element is usually mounted by itself below the wing of the aircraft concerned, but if not in such a protected position it needs a radiation shield. The bare contacts on the end of the elements need protecting from moisture. The element is connected by uncompensated leads to the control box which is, in effect, an unbalanced Wheatstone's bridge. Upon the box there are adjustments for the resistance of the leads and for the battery voltage. There are now three types of indicator. The one most commonly in use is an edgewise type. The scale has a range of +20°C to -60°C in 2°C divisions in a length of 1¾ inches, and beside the scale is a long mirror which is necessary to eliminate parallax error. The modification referred to in the introduction consists of substituting flat dial indicators for the edgewise type.

The new indicator is made in two versions, one reading from +20°C to -60°C in 2°C divisions, and the other reading twice over the range, once from -70°C to -10°C and once from -20°C to +40°C with 1°C divisions, the changeover being effected by a switch on the control box (which is of a slightly modified and more convenient design). The scale length is 2½ inches for the single reading type and 3 inches for the double reading type.

Photographs of the various modifications of this thermometer are appended.

2.2. General properties. The open element has a speed correction of 70% full adiabatic and this is not much affected by fitting a wide radiation shield; if, however, the back of the shield is partly blocked up by the mounting or by the moisture shield for the rear contacts on the element, then the speed correction may increase as far as 85% full adiabatic.

The lag of the thermometer is smaller than can be detected by methods in flight.

The accuracy is governed by the accuracy possible in the reading of the indicator and the accuracy of the calibration. A competent observer can read to a ¼ division on either of the flat dial types ($\frac{1}{2}$°C and $\frac{1}{4}$°C for the single and double reading types respectively); with the edgewise indicator an accuracy of $\frac{1}{2}$°C is possible if the reflection of the needle can be seen, but if the lighting is bad, or if the indicator is badly mounted, or if the needle is at the end of the scale, parallax errors may be as high as 3°C.

Calibration is now made at Boscombe Down on the aircraft. The resistance of the element at 0°C is found in the laboratory, and then the reading of the dial is taken for each of a series of settings of a standard resistance box, placed across the contacts to which the element is connected. This is a quick and efficient method of ensuring the constancy of calibration and the general serviceability of the instrument, and for the meteorological aircraft at Boscombe Down it is normally repeated about once a week.

Experiments at R.A.E. have shown that the indicators upon these thermometers can resonate to frequencies likely to be met with on flight. However, very little trouble has been met with at A.& A.E.E. despite considerable experience.

Trouble has lately arisen with the flat dial indicators due to sticking of the coil on the field magnet. This is due to the fact that the coil mounting bearing is very loose and in the flat dial indicators the coils are mounted sideways, whereas in the curved dial indicators the coil is mounted upright. This objection should be very easily removed by a more careful mounting of the coil; steps are being taken to have this done.

As can be seen above there are many minor troubles which can arise with these thermometers, but with constant serviceing the instrument is very useful and convenient for routine meteorological ascents.

3. The Negretti and Zambra Nitrogen filled capillary thermometer.

3.1. Description. The element of this thermometer is a metal tube about 12" long and $\frac{1}{2}$" diameter. This tube contains nitrogen under pressure. The element is usually mounted under the wing with its length along the span, protected by a streamlined radiation shield. Connection to the indicator is made by a fine copper capillary protected by a black composition sheath. The indicator is small and the needle, which is rather thick, revolves once for a range of +35°C to -60°C. The divisions are 2°C.

A photograph of the dial and element are appended to this report.

3.2. General properties. The speed correction upon these thermometers seems to vary rather randomly between 75% and 95% full adiabatic; it is believed that the correction is rather sensitive to the exact mounting of the radiation shield. 80% may be taken as a normal average. For high speed machines it is wise to make a determination for each mounting.

The lag of the thermometer is small but measurable, the decay from one temperature to another is approximately exponential with a $\frac{1}{2}$ period of 8 seconds. 30 seconds is therefore ample time for stabilisation.

The accuracy of the thermometer is not great. Accurate calibration in the laboratory has so far proved almost impossible due to the size of the element, and it is best calibrated by comparison in flight trials. Despite the crudity of the thermometer and the possibility of leakage, errors of more than 2°C are extremely rare. The thermometer, once calibrated, is very consistent and may be trusted to be accurate to 1°C or 2°C for considerable periods of time.

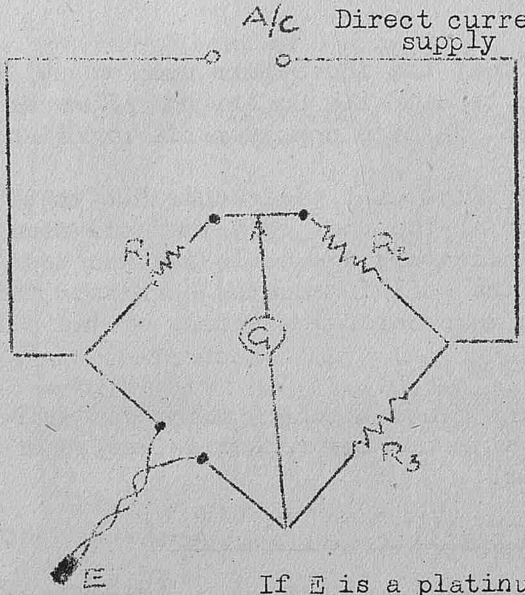
There is a possibility of parallax error which can be eliminated by suitable mounting or by an intelligent observer.

Since the thermometer is a constant volume pressure thermometer, there is a slight correction for external pressure ammounting to .17°C per 100 milli-bars or for normal purposes .05°C per 1000 ft.

The above properties do not make the thermometer suitable where high accuracy is required or as a permanent fitting to operational aircraft, but at this Establishment it serves the very useful purpose of indicating the approximate temperature consistently and thus showing the existence of a gross error on any of the more accurate and more sensitive thermometers in the aircraft.

4. Balanced bridge electrical resistance thermometers.

4.1. Description. Thermometers of this type have considerable possibilities mainly because it is easier to make a permanent resistance with negligible temperature coefficient than it is to make a milliammeter with the same properties. The only air thermometer of this type as yet in use is one developed by Mr. Brewer specifically for the routine measurement of air temperatures to an accuracy of about ½°F or ¼°C. Accuracy is sacrificed to compactness, the final complete instrument being small enough to mount easily in the gun sight mounting of a Spitfire. The circuit is as below:-



R_1 -)
) Standard
R_2 -) Constantin
) Resistances
R_3 -)

v Variable resistance

E Thermometer element

G Indicator

If E is a platinum resistance of about 100 Ω (The Cambridge Element has a resistance of 110 Ω at 0°C) and if the bridge has $R_2 = R_3$, then connected to a 12-volt battery supply G needs approximately the sensitivity of a milliameter. In the thermometer under discussion, (v) is a single turn of "Staybrite" steel wire wound upon a spool of about 3" diameter. A photograph of the instrument as mounted in a Meteorological Spitfire is appended to this report, the element used is that supplied for the Cambridge electrical resistance thermometers.

4.2. General Properties. The thermometer has so far shown practically no vices, the effect of the temperature coefficient of various parts of the apparatus is negligible, and the disadvantage of the sensitivity of the indicator is eliminated. As yet the only trouble met with in the thermometer is in the material and construction of the variable resistance. The scale is open, the likely parallax error is very small, and the accuracy is considerable although for the measurements required it is only read to the nearest ¼°C. The only disadvantage of the instrument is the fact that the observer has to adjust it to obtain a reading, but the pilots at the A.& A.E.E. consider that the open scale amply outweighs this disadvantage when compared with a Cambridge electrical thermometer.

This thermometer is designed for and is admirably suited for the routine measurement of upper air temperatures.

4.3. Further developments. It should be possible to produce an accurate and dependable standard thermometer on this principle.

5. R.A.E. Spirit in glass thermometer.

5.1. Description. This instrument is fully described in A.& A.E.E. Inst./37. It is a small L shaped spirit in glass thermometer with the bulb outside the aircraft and the stem inside. It is small and very robust. A photograph of this thermometer is appended to the report.

/5.2.

5.2 General properties. The objections raised in A.& A.E.E.Inst./37 to this thermometer are:-

(i) The large stem correction that can arise
(ii) The possibility of large parallax errors
(iii) The method of marking of the scale

In all it is believed that it is possible to produce a maximum error of as much as 5°C, but it is also stated that this error could be cut down to less than 2°C which would make the thermometers extremely useful as a standard fitting to operational aircraft.

6. N.P.L. Multi-junction thermocouple thermometer.

6.1 Description. The full description of this thermometer is given in A.& A.E.E. Inst.40. Briefly the thermometer is made up of six copper constantin couples mounted in series with a Turner micrcammeter (0-50 micro amperes). One set of junctions is mounted in a glass tube resting in melting ice. The other junctions are mounted in a pitot head. Two such pitot heads are mounted on the aircraft, one facing upstream and one facing down-stream; the down-stream facing pitot was designed to avoid icing troubles. Photographs of the microammeter and the pitot heads are appended.

6.2 General properties. The instrument has the following advantages :-

An accuracy of about 1/5th °C.
Full adiabatic speed correction to the forward facing pitot.
Simplicity and robustness of all components except the microammeter.
Ease of calibration (it is only necessary to check the microammeter upon each recalibration)
Negligible time lag

There are the following disadvantages :-

Temperature coefficient of the indicator
Extreme sensitivity of the indicator
The necessity of a reversing switch in the circuit
The necessity of renewing the ice on each trip

The above make the instrument suitable as a research standard instrument.

6.3 Further developments. In A.& A.E.E.Inst.40 it is recommended that:-

A suitable reversing switch and a switch for bringing the series resistance in and out should be included in the circuit.

A small thermometer should be mounted on the indicator to allow for changes of cabin temperature.

7. General discussion and conclusions.

The properties and uses of the various thermometers on the basis of the requirements set out earlier are summarised in Table I.

/7.1.

7.1. Condition (i) is fulfilled by the N.P.L. thermometer, an accurate strut thermometer and should be met by a suitably designed balanced bridge resistance thermometer. The strut thermometer is at the moment the most accurate and dependable, but suffers from the disadvantages of having a restricted range and being confined to multi-engined aircraft, or single engine biplanes.

7.2 Condition (ii) is well fulfilled by the double reading Cambridge Electrical thermometer, and the balanced bridge thermometer designed by Mr. Brewer; of these the second is definitely superior and fulfills nearly all demands.

7.3. Condition (iii) is not fulfilled by any of the thermometers mentioned in this report, but the R.A.E. spirit thermometer is a step in the right direction.

7.4 The Negretti and Zambra nitrogen filled capillary thermometer fulfills a valuable function as an installation for checking gross errors on routine meteorological ascents.

Circulation List.

C.R.D.	
D.C.R.D.	
D.T.D.	
D.D.T.D.	
D.O.R.	
D.D.R.D.A.	
D.D.R.D.T.	
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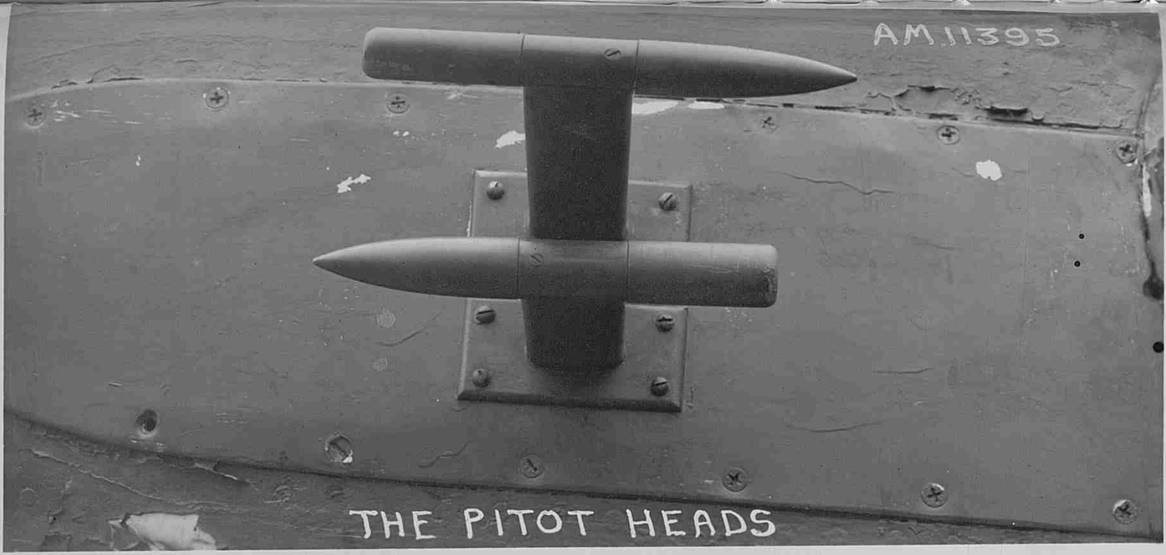
A.D.R.D.L.	4 copies
D.D.R.D.Inst.	2 copies (1 for Action)
R.D.Inst.	
D.D.S.R.2.	
T.D.Admin.	6 copies
R.T.P.2.	25 copies+ 1

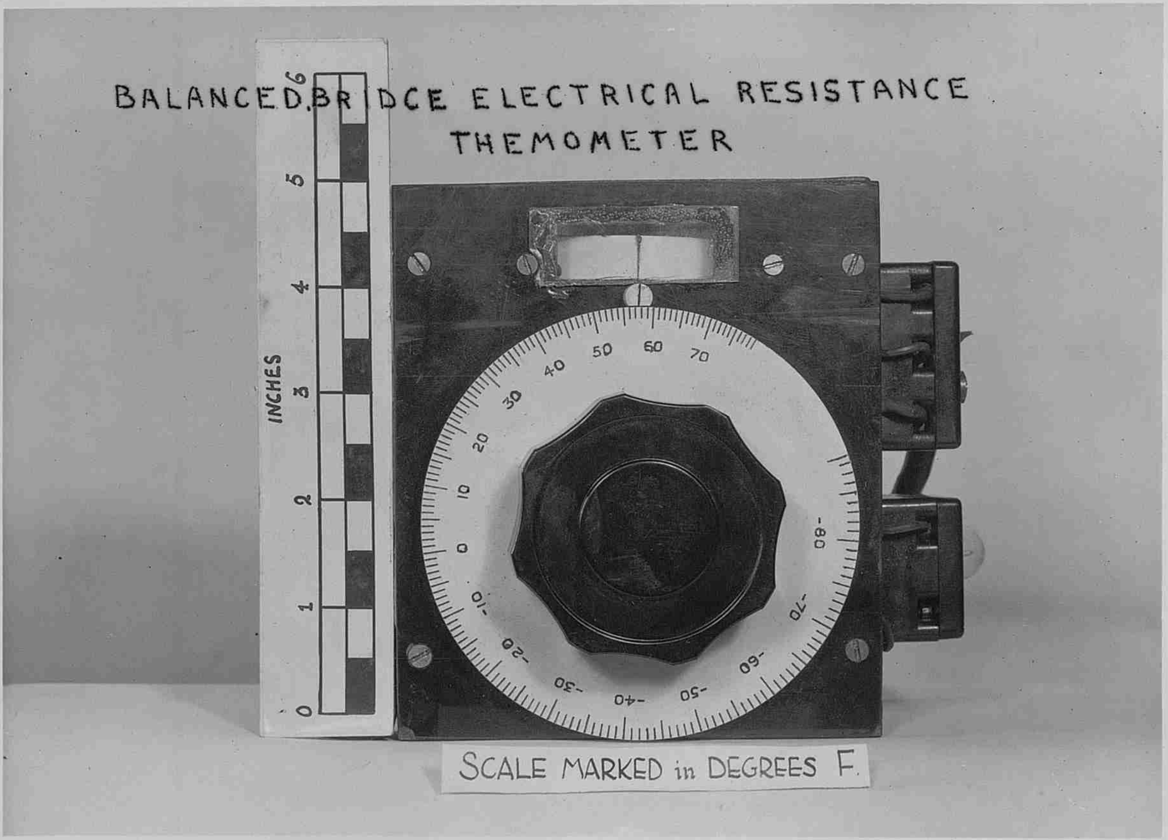
TABLE I.

Summary of properties of thermometers tested at A.& A.E.E.

Type of Thermometer		Max. possible accuracy with all corrections applied	Speed correction %age of full adiabatic	Possible parallax error	Simplicity and robustness	Time necessary for complete stabilisation	Ease and permanency of calibration	Other possible errors	Other possible disadvantages	Uses
N.P.L. Multi-junction thermocouple		1/10°C over restricted range	Upstream Pitot 100% Downstream pitot 75%	< ¼°C	Apparatus simple, indicator extremely sensitive	negligible	Easy to calibrate Permanency depends upon indicator	Temperature coefficient of indicator	Constant renewal of ice necessary. Some switching device necessary.	Fairly suitable for condition(i)
Meteorological strut Psychrometer		1/10th°C	Varies with ark	Very small	Simple and robust	2 - 3 mins.	Needs careful calibration but is permanent	None	Must be mounted outside aircraft. Scale at present only to -45°C. Possible ill effect on aircraft handling	Suitable for condition (i)
Cambridge Electrical Resistance	Flat dial	½°C or ¼°C.	70% for the free element	small	Not very simple apparatus. The dials are not very robust but ease of calibration nearly counteracts this.	Negligible	Calibrations very simple, can be done on the a/c. Permanency depends upon indicator	Battery error has suitable adjustment Temp.coefficient of the leads & the indicator does not give trouble.	Necessity for battery voltage adjustment. Possibility of resonance and sticking of the indicator	Both suitable for condition (ii)
	Curved dial	½°C.		up to 3°C.						
Balanced bridge Resistance		¼°C	70% for the free element	Very small	Not very simple Robust	Negligible	Calibration very simple, can be done on the aircraft.	None	Necessity of adjustment to get a reading	Very suitable for condition(ii)
Negretti and Zambra nitrogen filled capillary		1°C	Varies 75% - 95%	Up to 3°C.	Very simple. Requires great care to be taken that capillary tube is undamaged.	30 secs.	Extremely difficult to calibrate but appears to be fairly permanent	A correction for altitude is necessary.	Crudeness of indicator	As a crude meteorological thermometer fo indicating gross error.
R.A.E. Spirit in glass.		2°C.	65%	Up to 3°C.	Very simple and very robust	About ½ to 1 min.	Troublesome to calibrate but should (?) be permanent.	Stem correction is very large	Crudeness of scale.	With modifications may b suitable for condition(iii)



THE PITOT HEADS


INCHES
0
1
2
3
4
5
6
0
10
20
30
40
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60
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-10
-20
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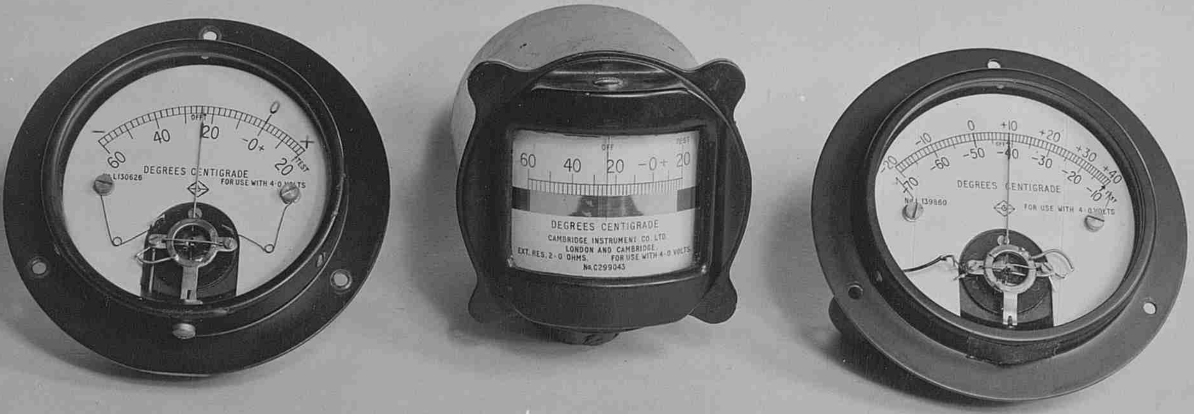
-80

BALANCED BRIDGE ELECTRICAL RESISTANCE THEMOMETER

SCALE MARKED in DEGREES F.

AM.11395.


60
40
OFF
20
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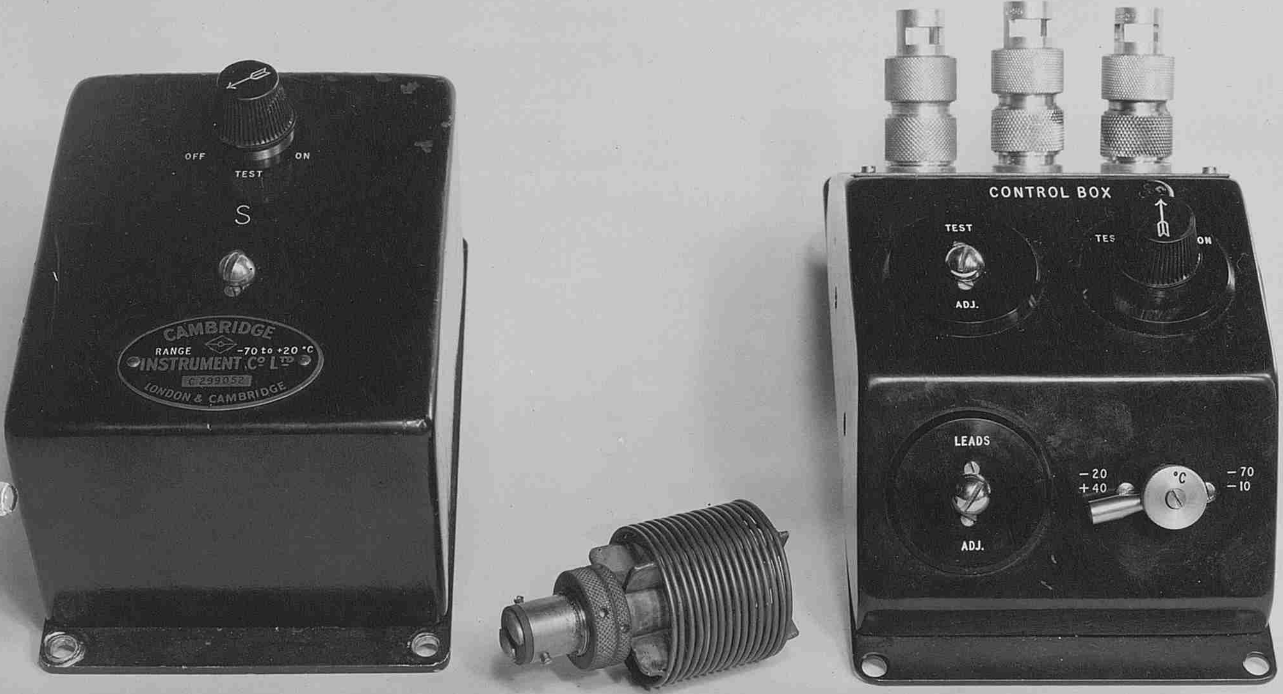
TEST
DEGREES CENTIGRADE
Nº L130626
FOR USE WITH 4·0 VOLTS
CAMBRIDGE INSTRUMENT CO. LTD.
LONDON AND CAMBRIDGE.
EXT. RES. 2·0 OHMS.
No. C299043
-70
-10
-60
-50
-40
-30
-20
+10
+20
+30
+40
Nº L139860

VARIOUS DIALS FITTED TO CAMBRIDGE
ELECTRICAL RESISTANCE THEMOMETERS


OFF
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S
CAMBRIDGE
RANGE -70 to +20 °C
INSTRUMENT Cº Lᵀᴰ
C 299052
LONDON & CAMBRIDGE
CONTROL BOX
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ADJ.
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LEADS
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-20
+40
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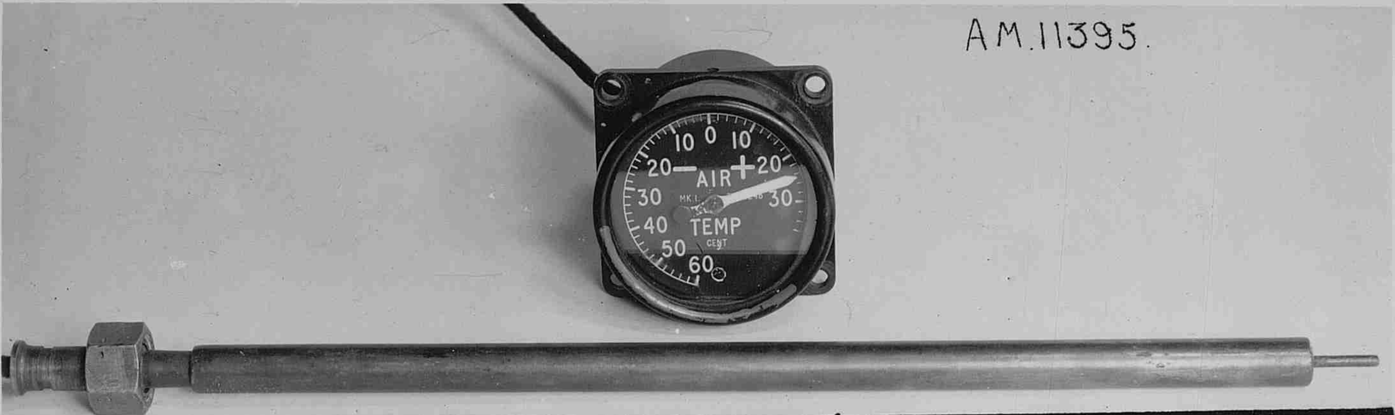
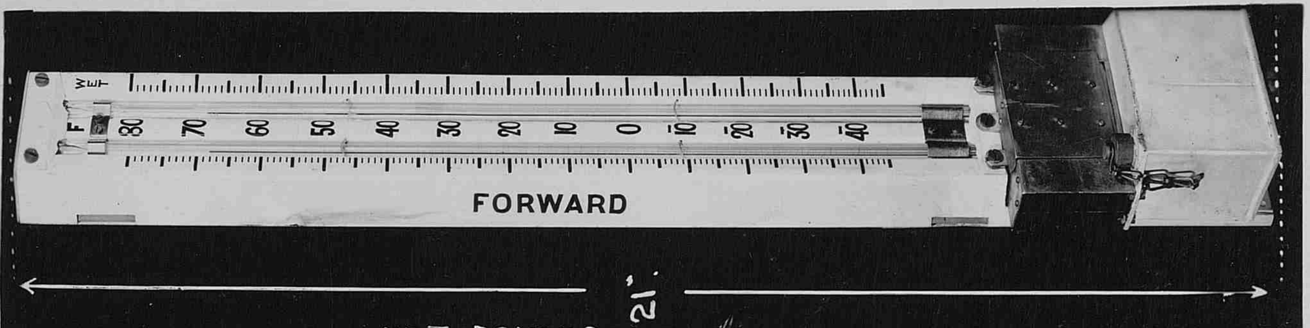
THE TWO TYPES OF JUNCTION BOX AND THE ELEMENT FITTED TO
CAMBRIDGE ELECTRICAL THEMOMETERS

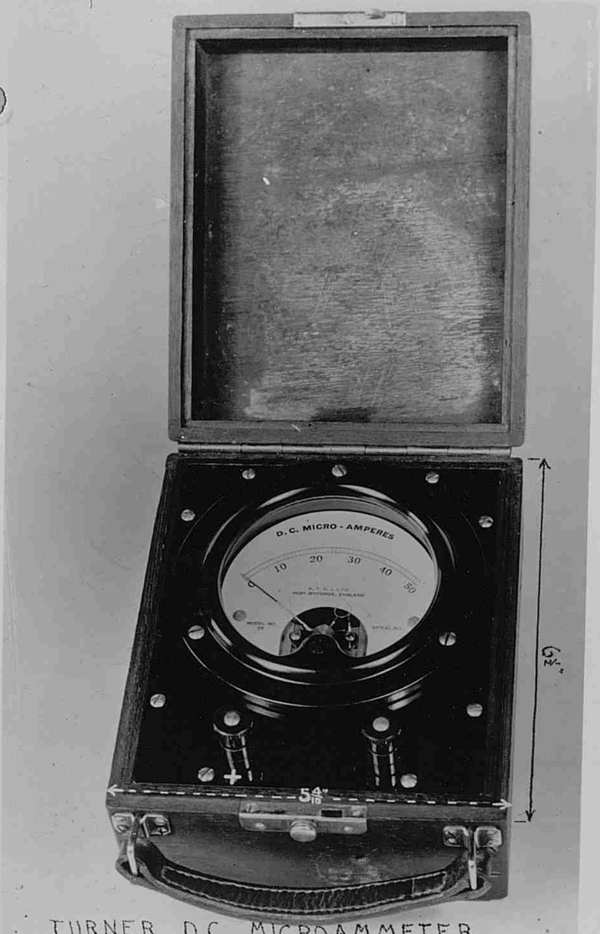
AM.11395.


AIR
TEMP
CENT
MK I
10 0 10
20 20
30 30
40
50
60

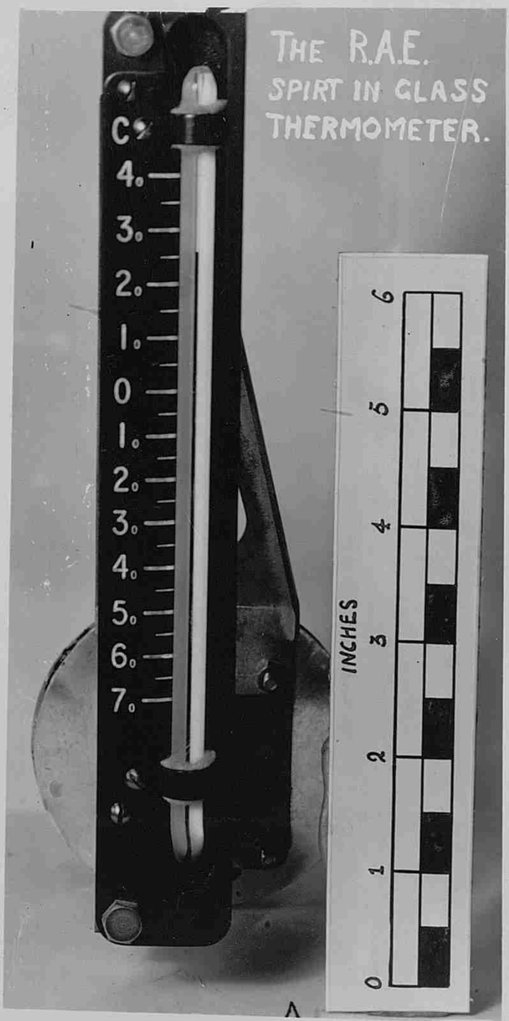
NEGRETTI AND ZAMBRA NITROGEN FILLED CAPILLARY THERMOMETER


WET
FORWARD
80 70 60 50 40 30 20 10 0 10 20 30 40
21"

MK V PSYCHROMETER


D.C. MICRO-AMPERES
10 20 30 40 50
+
-
$5\frac{4}{16}$"
$6\frac{1}{2}$"

TURNER D.C. MICROAMMETER


C°
4. 3. 2. 1. 0 1. 2. 3. 4. 5. 6. 7.
INCHES
0 1 2 3 4 5 6
A

THE R.A.E. SPIRT IN GLASS THERMOMETER.

U. S. ARMY AIR FORCES
WEATHER DIVISION
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M.R.P. 122
22nd July, 1943.

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Measurement of Turbulence in the Free Atmosphere.

(Comments on M.R.P. 64)
by W. Swinbank.

In connection with the note on "Measurement of Turbulence in the Free Atmosphere", M.R.P.64, by Mr. C.S.Durst, it may be worthwhile to mention the following points:

1. It is stated on page 4 that "it is remarkable that on several occasions the curves of variation of swing with height are almost exactly the mirror of the curves of the expression

$$\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2 - \frac{g}{T}\left(\frac{\partial T}{\partial z} + \Gamma\right)."$$

It should be remembered, however, that this expression of Richardson's is a measure not of the degree of turbulence, but of the tendency to change of turbulence. Now if the degree of turbulence is very great, the most probable change from this state will be one of decrease; and if the turbulent motion is very small, increase can be the only change.

Therefore it is to be expected that when the instrument has a wide range of swing, that is the degree of turbulence is great, the turbulent tendency shown by Richardson's Criterion, if this is correct, should be for decrease. And when the degree of turbulence is small, the only possible change is one of increase.

If the foregoing reasoning is correct, the observed fact that the two curves are images of one another is rather to be expected than remarkable.

2. The fact that the balloon shows greater swing in one region than in another does not necessarily mean that the degree of turbulence, or the total turbulent energy, is greater in the first region. It does mean that in this region there is greater energy associated with eddies whose size is comparable with that of the balloon system. Only when the eddies in the region are of the same order of magnitude as the system itself will the balloon be disturbed. If the eddies are very small they will not perturb the balloon; and similarly if they are very great the disturbance will be very gradual, and should not produce oscillations. It may be that in regions where the balloon is comparatively undisturbed there is considerable eddying energy, but not in that part of the spectrum of eddy size energy corresponding to the balloon size.

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M.R.P. 305-

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Date Initials d.T.

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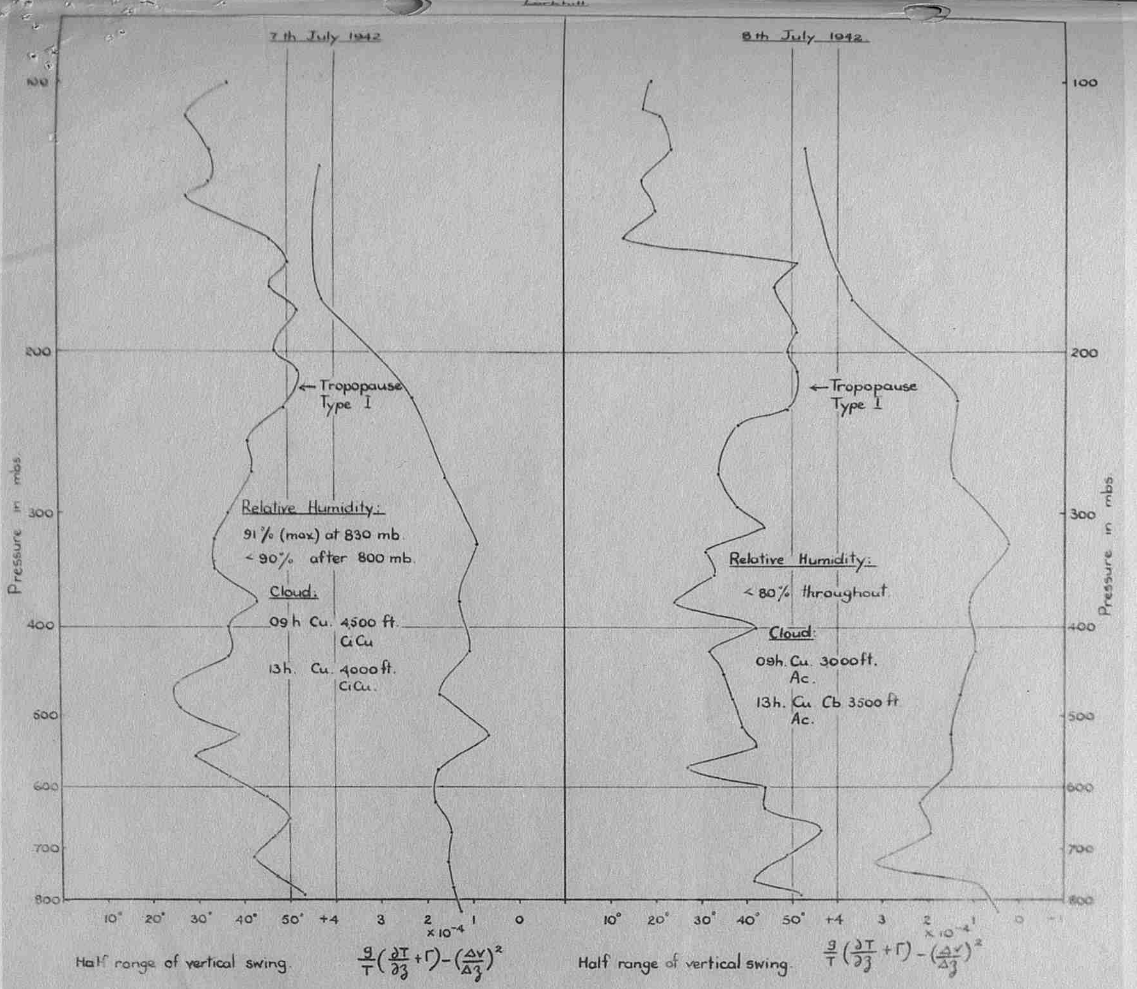
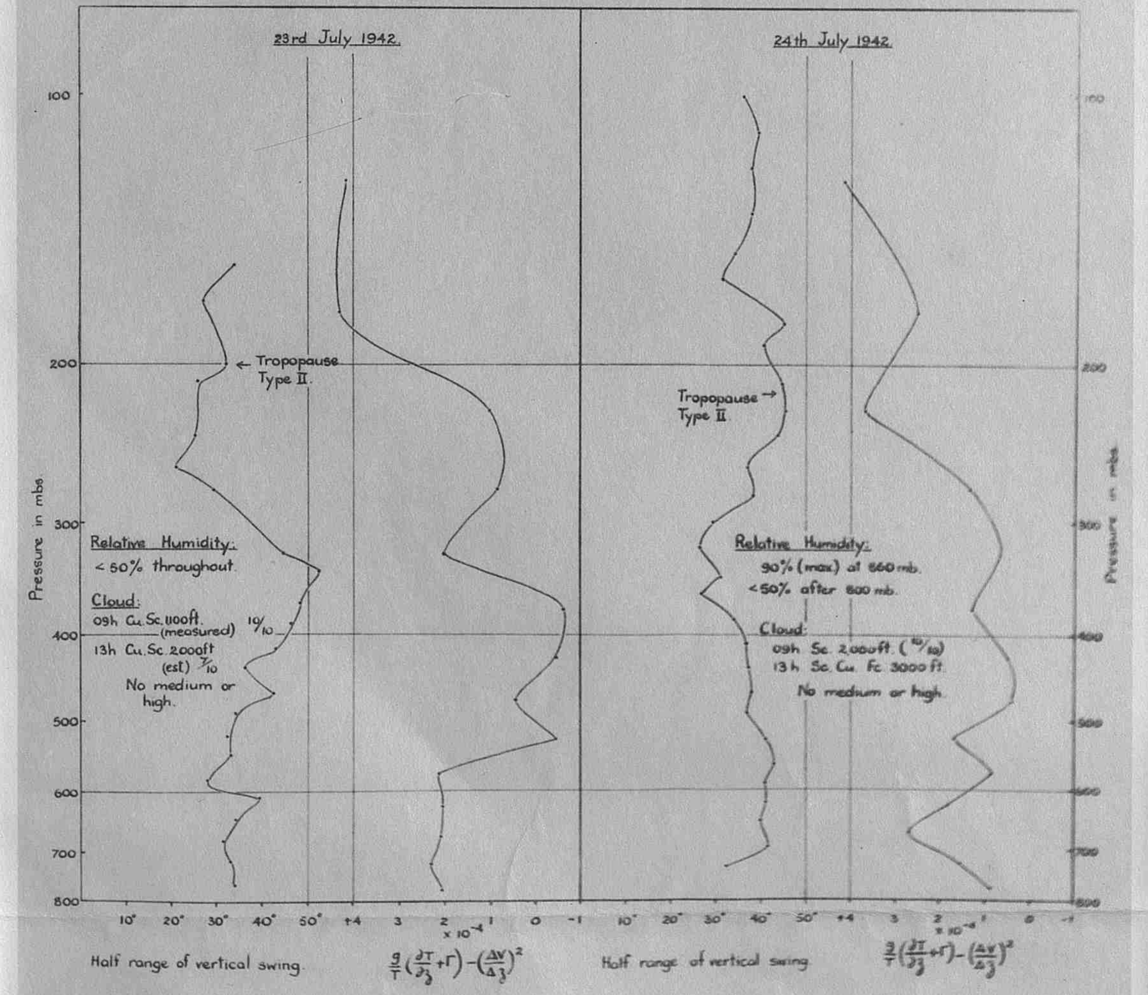
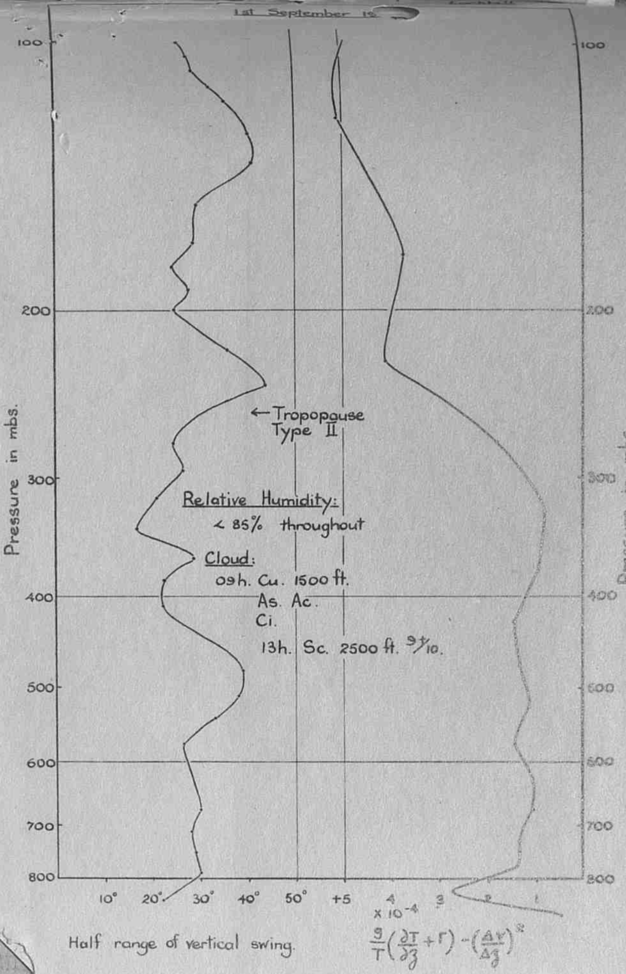

7th July 1942
8th July 1942
100
200
300
400
500
600
700
800
Pressure in mbs
←Tropopause Type I
Relative Humidity:
91% (max) at 830 mb.
< 90% after 800 mb.
Cloud:
09h Cu. 4500 ft. CiCu
13h. Cu. 4000 ft. CiCu.
Relative Humidity:
< 80% throughout.
Cloud:
09h. Cu. 3000 ft. Ac.
13h. Cu. Cb 3500 ft Ac.
10° 20° 30° 40° 50° +4 3 2 1 0 -1
x 10⁻⁴
Half range of vertical swing.
$\frac{g}{T}\left(\frac{\partial T}{\partial z}+\Gamma\right)-\left(\frac{\Delta v}{\Delta z}\right)^2$

Larkhill.


23rd July 1942.
24th July 1942.
100
200
300
400
500
600
700
800
Pressure in mbs
← Tropopause Type II.
Tropopause → Type II.
Relative Humidity:
< 50% throughout.
Cloud:
09h Cu. Sc. 1100 ft. (measured) 10/10
13h Cu. Sc. 2000 ft (est.) 7/10
No medium or high.
Relative Humidity:
90% (max) at 860 mb.
< 50% after 800 mb.
Cloud:
09h Sc. 2000 ft. (10/10)
13h Sc. Cu. Fc 3000 ft.
No medium or high.
10° 20° 30° 40° 50° +4 3 2 1 0 -1
x 10⁻⁴
Half range of vertical swing.
$\frac{g}{T}\left(\frac{\partial T}{\partial z}+\Gamma\right)-\left(\frac{\Delta v}{\Delta z}\right)^2$

1st September 19



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MRP 3/05

A METHOD OF EXTRAPOLATING THE PRESSURE DISTRIBUTION ALOFT BY THE AID OF REPRESENTATIVE SURFACE OBSERVATIONS.

by T.H. Martin.

1. Introduction.

In analysing upper air charts it is frequently necessary to extend the analyses (particularly the pressure contour lines) into regions from which no upper air observations are available.

In extending the contours into such regions the synoptic meteorologist will be guided by the following factors:-

(a) The isobaric and air mass analysis of the surface weather chart,
(b) Actual surface observations.

An extension of the contours according to (a) above will often render results that are qualitatively satisfactory. Particularly when the surface charts indicate the presence of well-defined types of air masses, the normal thickness of isobaric sheets in the various types of air masses can be used for estimating the height of (say) the 500 mb. surface above the 1000 mb. surface, the contours of the latter surface being obtained from the isobars of the surface chart. The method can at best lead to qualitative results, and the errors are likely to be appreciable when the structure of the air masses deviates from idealized types, and also when the history of the air mass is doubtful.

A method of computing the heights of the 750 and 500 mb. surfaces by the aid of representative surface observations was introduced into the daily routine of the Upper Air Unit by Dr.Petterssen, who asked the writer to determine the accuracy of the method by means of such observations as might gradually become available.

The method consists in selecting points on the surface weather charts where there is clear evidence of the presence of convective phenomena in deep layers of air, and computing the heights of the 750 and 500 mb. surfaces using the surface pressure (p_0) and temperature (T_0) and assuming that the mean lapse rate in the air column is equal to the saturated adiabatic rate.

This method, which has been in use since January 1943, has rendered results that appear to be quite satisfactory. To obtain an estimate of the accuracy of such extrapolation, the method was applied in 281 cases where the extrapolated values could be compared with actually observed heights. The results of these comparisons are given in para 4.

2. Representative Temperatures and Lapse-rates.

The air temperature near the earth's surface is said to be representative when the lapse rate is uniform throughout a deep layer of air above the surface. Representative surface air temperature may occur under many circumstances, and in particular when deep convective clouds are present. In cases of instability, the lapse rate is not only fairly uniform in each particular case, but it also tends to be constant from one case to another. On the average, the lapse rate under a convective cloud is not far removed from the dry-adiabatic; within the cloud it is not far removed from the saturated adiabatic, and above the cloud it is usually somewhat less than the saturated adiabatic (except when the cloud is growing rapidly).

The lapse rate of temperature outside the cloud may differ locally from that within the cloud, but the mean temperature of the air column outside the cloud cannot differ appreciably from the mean temperature of the column through the cloud, because otherwise very strong localised pressure gradients would occur in the upper atmosphere around the clouds. As a first approximation it may, therefor be assumed that the mean lapse rate in convective air is approximately saturated-adiabatic, and if the surface air temperature is representative, the mean temperature of the air column can be represented by the saturated adiabatic through the point on the Tephigram that represents the surface observation (p_o, T_o). At each particular level the actual temperature may differ from that indicated by this saturated adiabatic, but the mean temperature of the air column ought to be fairly close to the mean temperature indicated by the saturated adiabatic through the point (p_o, T_o).

Since the height of a pressure surface (p) above a pressure surface (p_o) depends only upon the mean temperature of the air column, an approximate value of the height can be thus obtained.

Since strong winds tend to stir the air column the surface-air temperatures tend to be representative in strong winds. However, in the absence of convective phenomena there is no guarantee that the lapse rate in strong wind is near a critical value. Nevertheless, it is of interest to determine whether or not the lapse rates in strong winds are such that extrapolations are permissible.

3. Criteria.

The following criteria, which are ascertainable from the synoptic weather map, have been used in selecting suitable points for extrapolating heights:-

Cumulus congestus	△
Cumulo-nimbus	Ξ
Rain showers	▽
Hail	⩡
Thunderstorms	Ꞧ
Wind of force 5 or more without convection clouds.	⫻—

Cumulus humilis and similar clouds were excluded since they often occur in relatively shallow layers.

4. Accuracy.

Table I shows the mean errors in the extrapolated heights for the 750 and 500 mb. surfaces for various regions.

TABLE I.

Mean Errors in Extrapolated Heights (feet)

Region	△ Ξ ▽ ⩡ Ꞧ		⫻—	
	750 mb.	500 mb.	750 mb.	500 mb.
British Isles	+17	+26	-49	-240
Atlantic	+35	+32	-26	-140
Iceland	+25	-15	-43	-250
Mediterranean	+19	+50	-	-

It will be seen that, on the average, the heights of the 750 and 500 mb. surfaces can be obtained from the surface observations with sufficient accuracy when there is clear evidence of instability phenomena. It will also be seen that the mean errors in the extrapolations are appreciable if strong wind is used as a criterion.

Furthermore, the fact that the mean errors are positive in the instability group indicates that, on the average, the actual lapse rate in the air column is steeper than the saturated adiabatic. The reverse is true when there are strong winds without instability clouds.

It is worthy of note that the mean errors in the instability group are not appreciably larger in the 500 mb. column than they are in the 750 mb. column.

Table II shows the percentage frequency distribution of errors.

TABLE II.

Percentage Frequency of Errors in Extrapolated Heights.

± Error (Feet)	⌓ ⌓ $\dot{\triangledown}$ $\overset{\triangle}{\triangledown}$ ⚡		⌙⌙⌙—	
	750 mb.	500 mb.	750 mb.	500 mb.
0-50	68	25	65	8
50-100	27	16	18	14
100-150	4	25	14	14
150-200	1	15	3	11
200-250		7		6
250-300		7		11
300-350		4		11
350-400		1		6
400-450				5
450-500				1
500-550				4
550-600				5
600-650				3
650-700				0
700-750				1

This table shows:

(a) that with instability as a criterion the height of the 750 mb. surface can be computed with a 95 per cent chance of being correct within 100 ft.

(b) that with strong wind as a criterion the height of the 750 mb. surface can be computed with an 83 per cent chance of being correct within 100 ft.

MRP 127

(c) that with instability as a criterion the height of the 500 mb. surface can be computed with an 81 per cent chance of being correct within 200 ft.

(d) that with strong wind as a criterion the chance for the height of the 500 mb. to be correct within 200 feet is only 47 per cent, and that the chances for very large errors are appreciable.

It will furthermore be seen that although the heights extrapolated with instability as a criterion may not be sufficiently accurate for use together with observed heights within a relatively dense network of stations, they are in the great majority of cases sufficiently accurate for obtaining the large-scale pressure distribution over areas where observations are lacking.

Although strong wind alone is not a sufficient criterion for extrapolating the height of the 500 mb. surface, strong wind together with instability cloud has usually been found to be a better criterion than convective clouds with light wind. Strong wind with cyclonically curved isobars appears to be a better criterion than strong wind with anticyclonic isobars, presumably on account of the subsidence that often develops in anticyclonic circulations.

Finally, it should be noted that instability as a criterion should not be used in connection with fronts, since the lapse rate in the frontal zone may vary considerably along the vertical.

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MRP 129
(Reprint)

LARGE SCALE TURBULENCE IN THE FREE ATMOSPHERE

AIR WEATHER SERVICE
TECHNICAL REFERENCE FILES
APR 23 1952

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1ST IND., DATED 23 APRIL 1947

CHIEF OF NAVAL OPERATIONS

AEROLOGY SECTION

WASHINGTON, D. C.

FEBRUARY, 1945

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Aerology Bull # 9-46
13 Apr. 1946

MRP 129

16 Aug 1943

ON THE APPLICATION OF THE RICHARDSON CRITERION TO LARGE SCALE TURBULENCE IN THE FREE ATMOSPHERE

By

Sverre Petterssen and W.C. Swinbank

From a Publication by
Air Ministry
Meteorological Research Committee

August, 1943

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ON THE APPLICATION OF THE RICHARDSON CRITERION TO LARGE SCALE TURBULENCE IN THE FREE ATMOSPHERE

1. Introduction.

A brief report* on an investigation into the relationship between the wind shear and the static stability of the air in the layer immediately below the tropopause was presented to the M.R.C. in October, 1942. In the course of the ensuing discussion, members of the M.R.C. suggested that the investigation should be extended to cover the entire troposphere. This has been done particularly with a view to finding empirical relationships between stability and wind shear and comparing these with the results obtained by applying the Richardson criterion to actual observations. A further demand for investigating this problem arose in connection with an investigation of the turbulence conditions that are obtained at the tropopause.

In the present investigation use has been made of the reports from the British radiosonde stations for the period December, 1942 to March, 1943. The accuracy of the Radar Direction Finding method of determining the wind in the free atmosphere depends to a large extent upon the geometry of each case, and a tolerance figure is allotted to each observation assessing the accuracy. Only reports of the highest degree of accuracy (i.e. vector errors less than 5 km/hour) have been used in this investigation. In all, 1556 simultaneous observations of lapse rate and wind shear have been investigated.

The observations have been grouped according to levels, and the results are illustrated in 12 diagrams in which a lapse rate is the abscissa and wind shear the ordinate. The diagrams contain also the limiting curve between decrease and increase in turbulence as computed from Richardson's criterion. Since wind observations are made at intervals of 50 millibars, the diagrams represent the observed relationship between the mean wind shear and the mean lapse rate through layers of 50 mb. depth. The limiting curve in these diagrams would then represent the limit between the conditions that determine decrease and increase in the large scale turbulence of the atmosphere.

The diagrams show that, with very few exceptions, for any given degree of stability the observed wind shear is less than that required by Richardson's theory to cause the turbulence to increase.

If the assumptions underlying the theory were correct and applicable to large scale turbulence, the diagrams would indicate that large scale turbulence almost invariably tends to decrease. Since this result appears improbable, it is shown that agreement between theory and observation can be obtained by dropping one of the assumptions made by Richardson; namely, that the coefficient of eddy transfer of heat is equal to the coefficient of eddy transfer of momentum. The diagrams then lead to the conclusion that the former coefficient is smaller than the latter, the ratio between them being about 0.65.

2. Brief Summary of Richardson's Theory.

The role played by the eddy stresses that are brought into being when the motion of an incompressible fluid ceases to be laminar, and becomes turbulent, was first discussed by Reynolds,** who assumed that the work done by the eddy stresses is consumed in two ways, viz.

(a) in increasing the kinetic energy of the turbulent motion, and

(b) in increasing the intrinsic heat energy of the fluid by dissipation due to its molecular viscosity.

Reynolds showed that the kinetic energy of eddying will increase when the work done exceeds the heat energy thus created, and furthermore, that the rate of loss due to molecular viscosity becomes greater, and, therefore, the rate of gain in eddying energy less with increasing eddying velocity and decreasing size of eddies.

L. F. Richardson*** extended the work of Reynolds to the case of a compressible fluid, and, in particular, to apply to the free atmosphere, where the eddies, in their vertical motion, must do work on account of expansion.

---

* M.R.P. 51

** Phil. Trans. A. t clxxxvi p. 123.

*** Proc. Roy Soc., A, Vol. 97, 1920.

Richardson considers a large block of air, and, applying the principle of conservation of energy, determines the gain in eddying energy by estimating the losses from all the other available forms, namely:

(1) the intrinsic heat energy.

(2) the gravitational energy.

(3) the kinetic energy of the mean (i.e. non-turbulent) motion.

(4) the energy conducted to the block from the underlying surface.

(5) the energy due to radiation exchanges.

(6) the work done by pressure forces at the vertical boundary surfaces of the block.

By choosing the block sufficiently large, Richardson shows that (6) is negligible as against the energy derived from other sources, and by assuming that the temperature distribution is known, he is able to eliminate the effects of radiation. The sources (1) and (2) above can be treated as one source, which is hereafter referred to as heat energy. The loss of energy due to molecular viscosity is shown to be negligible except, perhaps, for very small eddies. Thus, if the diameter of the eddy were 10 meters, the loss of energy through this agency would amount to only 20 percent in 24 hours.

Richardson next considers the mixing by the eddies of portions of air at different temperatures, and discusses two types of activity, which determine the supply and removal of energy to and from the eddies. One of these, which is called the boundary activity, is shown to determine the rate at which heat is supplied to, or removed from the underlying surface. The second, which is called the body activity, determines whether or not heat energy is being converted into eddying energy.

From the equations relating these two activities, Richardson derives an expression for the rate at which heat energy is transformed into eddying energy, and shows that it depends upon the upward gradient of entropy per mass of the air. Finally, he combines the energy supplied by this means with that supplied by the mean wind, and arrives at the following equation:

$$\frac{\partial E'}{\partial t} - \frac{2\mu_c \pi^2 E'}{\rho \ell^2} = \iiint \mu \left[\left(\frac{\partial \bar{V}_x}{\partial h} \right)^2 + \left(\frac{\partial \bar{V}_y}{\partial h} \right)^2 \right] dh\, dA$$

$$- \iiint \frac{Cg}{\gamma_p} \frac{\partial \sigma}{\partial h}\, dh\, dA$$

Here the symbols have the following meanings:

E' = kinetic energy of eddying

μ_c = coefficient of molecular viscosity

ρ = density of the air

ℓ = a linear measure of the size of the eddies

μ = coefficient of eddy viscosity

C = coefficient of eddy conductivity

h = height

$\bar{V}_x, \bar{V}_y$ = mean horizontal wind velocity along perpendicular axes

A = cross section area of atmospheric block

g = acceleration of gravity

γ_p = specific heat of air at constant pressure

σ = entropy per mass of the air.

Following G. I. Taylor*, Richardson now assumes that the coefficient of eddy viscosity (i.e. the coefficient of eddy diffusivity of momentum) is equal to the eddy conductivity of heat; in other words

$$\mu = C$$

The small term due to molecular viscosity is neglected and equation (1) reduces to

$$\frac{\partial E'}{\partial t} = \iiint c\left\{\left(\frac{\partial \bar{V}_x}{\partial h}\right)^2+\left(\frac{\partial \bar{V}_y}{\partial h}\right)^2-\frac{g}{\gamma_p}\,\frac{\partial \sigma}{\partial h}\right\}dh\,dA$$

This means that the sign of the rate of change of the kinetic energy of eddying (E') of the block depends upon the sign of the integrand

$$\left(\frac{\partial \bar{V}_x}{\partial h}\right)^2+\left(\frac{\partial \bar{V}_y}{\partial h}\right)^2-\frac{g}{\gamma_p}\,\frac{\partial \sigma}{\partial h}$$

The last term in this expression can be transformed so as to contain the vertical gradient of temperature, and becomes

$$\frac{g}{T}\left(\frac{\partial T}{\partial z}+\Gamma\right)$$

where T is temperature and Γ is the dry-adiabatic lapse rate. In this form, Richardson's criterion states that the kinetic energy of the eddying within a layer will increase or decrease according to whether

$$\left(\frac{\partial \bar{V}_x}{\partial h}\right)^2+\left(\frac{\partial \bar{V}_y}{\partial h}\right)^2 \gtrless \frac{g}{T}\left(\frac{\partial T}{\partial z}+\Gamma\right)$$

The assumptions underlying this criterion refer partly to the nature of the energy transformations and partly to the numerical relation between the coefficients for eddy transfer of momentum and heat. While the assumptions concerning the energy transformations seem entirely justifiable, there is very little evidence in support of the view that the coefficient of eddy transfer of momentum is equal to the coefficient for eddy transfer of heat. If the one differs from the other, the right hand side of the above inequality would contain a factor differing from unity.

Equating the shear term to the stability term in the above inequality, we obtain the equation for the critical curve separating cases of increase from cases of decrease in turbulence. This curve has been drawn in the diagrams that will now be discussed.

3. Constructions of Frequency Diagrams.

In the diagrams appended the abscissa indicates the decrease in temperature (°F) from the base to the top of a layer of air of 50 mb. Thus, the abscissa indicates lapse rate of temperature with regard to pressure, zero indicating isothermal conditions, negative values temperature increase, and positive values temperature decrease along the vertical.

The ordinate indicates the magnitude of the vector difference between the wind at the base and at the top of the layer, expressed in km/hour.

The parabolic curve in each diagram indicates the Richardson criterion as a function of lapse rate and wind shear. The curve intersects the abscissa in the point corresponding to the dry-adiabatic lapse rate. Since the Richardson criterion is expressed in terms of shear and lapse rate with regard to height whereas the observations give similar quantities with regard to pressure, the critical curve has been computed with the conversion from height to pressure that holds for the standard international (I.C.A.N.) atmosphere. The errors due to this standard conversion are negligible in comparison to the observational errors.

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* Phil. Trans. A, 1914.

In view of the fact that it is difficult to assess the accuracy of the wind observations when the balloon is near the earth's surface the lowest layer included in the investigation is the one from 900 to 850 mb. The atmosphere from 900 to 300 mb. has been divided into layers of 50 mb., and a diagram made for each layer. On account of the frequent occurrence of the tropopause above the 300 mb. surface no diagrams have been compiled for the layers above this level. Each point in the diagrams represents a simultaneous observation of wind shear and lapse rate. A point in the diagram below the critical curve indicates that, according to the Richardson criterion, the turbulence tends to decrease, and a point above the curve indicates that it tends to increase.

The diagrams 1 - 12 show the frequency distribution of the points in relation to the critical curve, and diagram 13 refers to the behavior of an individual point.

4. Discussion of the Frequency Diagrams.

The points in the frequency diagrams may be referred to three categories according to whether they are below, on, or above the critical curve. The number of points belonging to each of these groups is given in Table I.

.TABLE I

Pressure Intervals (mb.)	Number of Points		
	Below	On	Above
900 - 850	103	2	12
850 - 800	112	1	4
800 - 750	131	0	6
750 - 700	139	0	1
700 - 650	143	0	6
650 - 600	135	0	3
600 - 550	142	0	3
550 - 500	133	0	3
500 - 450	105	0	7
450 - 400	104	1	4
400 - 350	100	0	6
350 - 300	145	0	5
SUM	1,492	4	60

It will be seen that less than 4 percent of the points occur above the curve, indicating that in less than 4 percent of the cases were the conditions such that the turbulence would increase; while it would tend to decrease in about 96 percent of all recorded cases.

Since the observations cover a period of four months, it is reasonable to assume that the mean increase in turbulence over the period investigated is nil. Therefore, if the time-trends of the variation in turbulence were symmetrical, or such that, on the average, the turbulence increased over as long time intervals as it decreased, an equal number of increases and decreases ought to be recorded, with the result that an equal number of points ought to be found above and below the critical curve.

On the other hand, the possibility exists that the increase in turbulence sets in suddenly while the stabilizing processes act slowly. If this were so, the chance for recording a case of increase would be less than that for recording a decrease, and the ratio of the number of recorded decreases to recorded increases would increase with the assymmetry of the process. The results shown in Table I could then be explained by assuming that the assymmetry is about 24 to 1.

Assuming a pronounced assymmetry in the process, it is evident that the intensity of the increase in turbulence must be larger than the intensity of the decrease, since the net change over an extended interval must be nil. The diagrams show, however, that the points suggesting an increase are very close to the critical curve while the center of gravity of the points suggesting a decrease is far removed from that curve. This, again, would indicate that, on the average, the intensity of the increase is less than the intensity of the decrease in turbulence. The diagrams 1 - 12 are, therefore, consistent neither with the assumption of symmetry nor with that of assymmetry.

The ideal solution would be to study the variation in turbulence of individual volumes of air. Only in very few cases were the wind direction and speed such that the air could be traced from one station to another. The results of the investigation of these cases, which will be described later, were not conclusive. On the other hand, analyses of balloon oscillations have shown rather conclusively that the periods of increasing and decreasing oscillations are of about the same length and there is no evidence to indicate any appreciable assymmetry in the processes. It seems therefore that the paradoxical results shown in the diagrams 1 - 12 must be due to either or all of the following factors.

(a) the application of the theory to turbulence on such a large scale that layers of 50 millibars are affected may not be permissible.

(b) condensation processes may contribute to the turbulent energy and these have not been accounted for in the theory.

(c) the assumptions underlying the theory might lead to numerical disproportions affecting the position of the critical curve.

5. The Scale Effect.

It is of interest to note that one of the assumptions upon which the theory is based is that the volume considered is sufficiently large for the work transmitted across its vertical walls to be negligible in comparison to the other forms of energy involved. This assumption, which seems highly plausible, does not indicate that appreciable discrepancies between theory and observation should be due to considering layers of air of 50 mb. depth.

If the depth of the layer considered were an important factor one would expect to find in the diagrams 1 - 12 a distortion increasing with decreasing pressure, since the thickness of the considered layers increases with decreasing pressure. The layer from 900 to 600 mb. is only about half as deep as is the layer from 600 to 300 mb. Nevertheless, it will be seen from Table I that these two layers contain about the same number of points above the critical curve. It appears therefore that the thickness of the layer does not seriously influence the results.

Strictly speaking, the application of the criterion to layers of 50 mb. depth is permissible only when both lapse rate and wind shear are uniform within each layer. Observations with multiple theodolites made at Larkhill show that, for the depths considered, the procedure is reasonable as far as wind shear is concerned, and examination of the Tephigrams indicates that in the large majority of cases, the same is true for the lapse rate. This is particularly true of the upper half of the troposphere.

6. Influence of Condensation.

If the air were saturated, the saturated-adiabatic rather than the dry-adiabatic lapse rate would enter as a parameter in the criterion for turbulence. In diagram 13, the upper curve is the critical curve for non-saturated conditions, and the lower one is the critical curve drawn on the assumption that the saturated adiabatic can replace the dry-adiabatic lapse rate in the

Richardson criterion when the air is saturated. This simple substitution of the saturated adiabatic may not be strictly correct, but it will suffice for the qualitative discussion that follows.

The points in diagram 13 represent consecutive observations at 6-hour intervals of the relation between lapse rate and shear during a frontal passage accompanied by condensation. It will be seen that when the saturated adiabatic replaces the dry-adiabatic, the critical curve is lowered and the points can more readily appear on the side of the curve where increase in turbulence is possible.

If the critical curve corresponding to saturation were drawn in the diagrams 1 - 12, about one-third of all points would be found above the curve and below the one that corresponds to non-saturated conditions.

The frequencies of relative humidity in excess of 97 percent at Larkhill during the period under investigation are:

900 mb.	17.7 percent.
750 mb.	8.1 percent.
500 mb.	3.4 percent.

Taking the average frequency of saturated cases to be 15 percent (as a high value) it will be seen that less than 5 percent of all the points in the diagrams 1 - 12 would be affected by condensation and there would therefore, all in all, be a large surplus of points indicating decrease in turbulence.

The importance of saturation can be tested also in the following manner:

At the low temperatures occurring in the uppermost layers (say above 400 mb.), the saturated adiabatic is almost identical with the dry-adiabatic, and only minute amounts of energy are provided through condensation processes. The critical curve corresponding to saturated conditions would not differ noticeably from that valid for non-saturated air. The large deficit of points above the critical line at these levels can, therefore, not be ascribed to the influence on the position of the critical curve caused by condensation.

7. The Coefficients of Eddy Transfer.

The critical curves in the diagram were drawn on the assumption that the coefficient of eddy transfer of heat is equal to that of eddy transfer of momentum. If the former were smaller than the latter, the critical curve would be lower, and the number of points above the critical curve would be larger than shown in the diagrams 1 - 12. Since there is very little evidence in support of the view that the ratio of the said coefficients is unity, we shall now endeavor to determine the ratio by assuming that, during the period under investigation, the net gain in turbulence was nil. If this is so, then the true critical curve must be represented by the curve through the mean ordinates of the points. By drawing the smoothed curves for the mean ordinates in each of the diagrams 1 - 12 the values shown in Table II for the ratio of the two coefficients result.

It will be seen from Table II that there is no significant variation with elevation, the ratio being about 0.65 throughout. For turbulence on the scale here considered, it may be assumed that the ratio of the coefficients is about 2/3.

With this value for the ratio of the coefficients, a new critical curve may be drawn in all the diagrams. Relative to this new critical curve 42 percent of the points will be found to indicate a tendency for increase and 58 percent of the points a tendency for decrease in turbulence. This assymmetry might suggest that on the average, the periods during which the turbulence increases are somewhat shorter than are those during which it decreases. With this assymmetry of about 4 to 6, the average intensity of the increase would stand as a ratio of the average intensity of the decrease as about 6 to 4.

TABLE II

Ratio of the Coefficient of Eddy Transfer of Heat to the Coefficient of Eddy Transfer of Momentum

Layer (millibar)	Ratio	Number of Observations
900 - 850	0.75	117
850 - 800	0.65	117
800 - 750	0.64	137
750 - 700	0.65	140
700 - 650	0.64	149
650 - 600	0.62	138
600 - 550	0.62	145
550 - 500	0.62	136
500 - 450	0.69	112
450 - 400	0.66	109
400 - 350	0.63	106
350 - 300	0.64	150
Mean for Layer 900 - 550	0.653	1043
Mean for Layer 550 - 300	0.650	513

8. Time Variations.

An attempt has been made to study the variations in the turbulence of individual samples of air. Only in 44 cases were the wind direction and speed such that the sample of air would move from one station to another during the time interval that elapsed between two consecutive observations. The frequency distribution of these cases is shown in Table III, in which + stands for "increasing" and — for "decreasing," and zero means "no change."

TABLE III

Lapse Rate	Shear			Sum
0	—	0	+	
—	9	1	5	15
0	4	4	5	13
+	8	0	8	16
Sum	21	5	18	44

It will be seen that a decrease in shear was as frequently associated with increasing as it was with decreasing lapse rate. The same appears to be true of increasing shear.

The apparent lack of correlation between the variations in wind shear and lapse rate seems to suggest that the time interval between the observations (i.e. 6 to 8 hours) is so large that the tendency at one instant is not appreciably determined by its value 6 to 8 hours earlier.

9. Conclusions.

The discussion in the foregoing paragraphs seems to indicate that relationships between wind shear and lapse rate in the free atmosphere are such that

(a) the Richardson criterion is applicable to large scale turbulence in the free atmosphere if the ratio of the eddy transfer of heat to that of eddy transfer of momentum is less than unity.

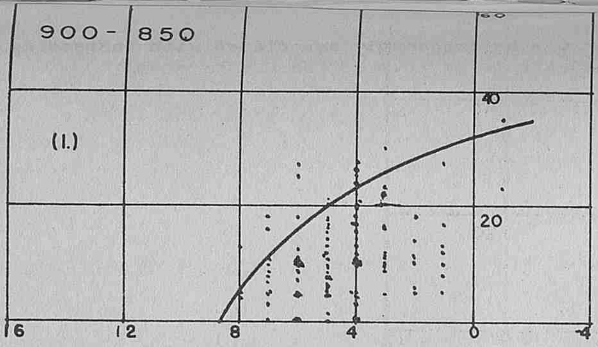
(b) the most probable value of this ratio is 0.65.

(c) above the friction layer this ratio is independent of elevation.

(d) condensation processes do not appear to have any appreciable direct effect upon the average conditions, although they may have an appreciable indirect effect through the lapse rates that they may create.

(e) the intervals through which the turbulence increases seems to be somewhat shorter than the interval through which it decreases, and that therefore the intensity of the increase is greater than the intensity of the decrease.

(f) the duration of the increase, or decrease, in turbulence in individual volumes of air is frequently less than 6 - 8 hours.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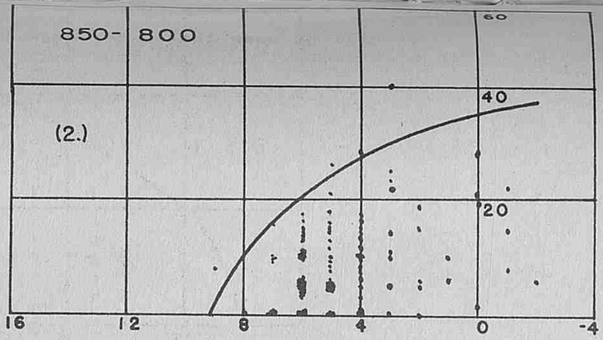
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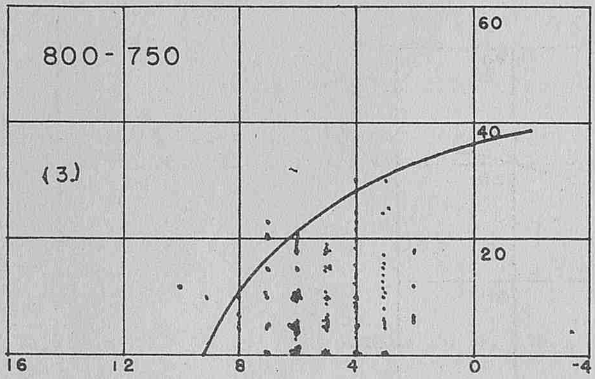
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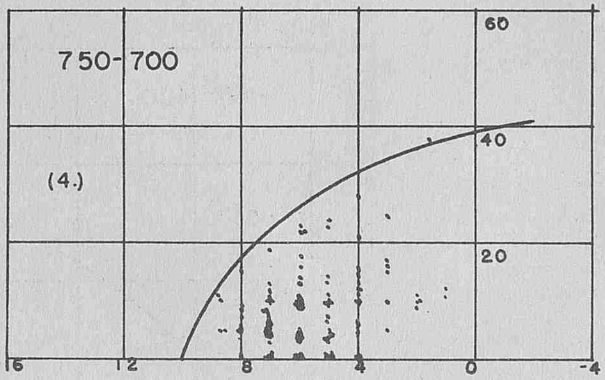
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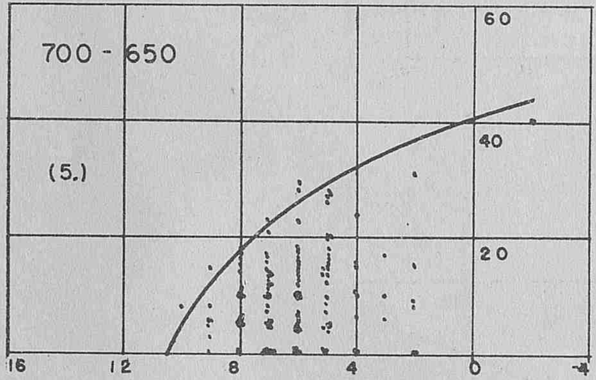
-4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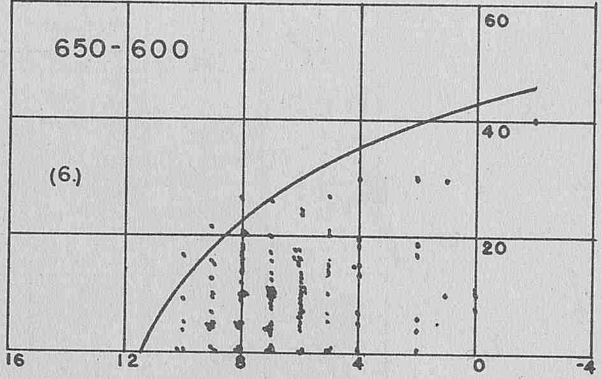
0
-4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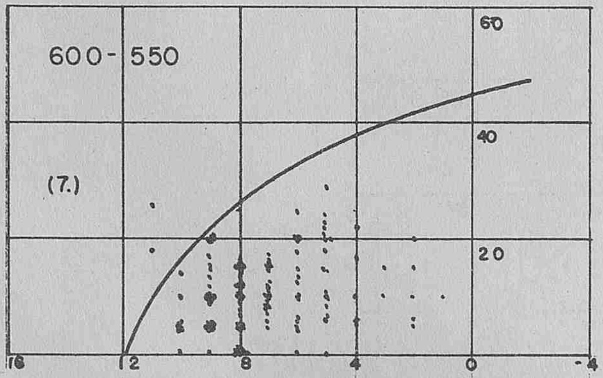
-4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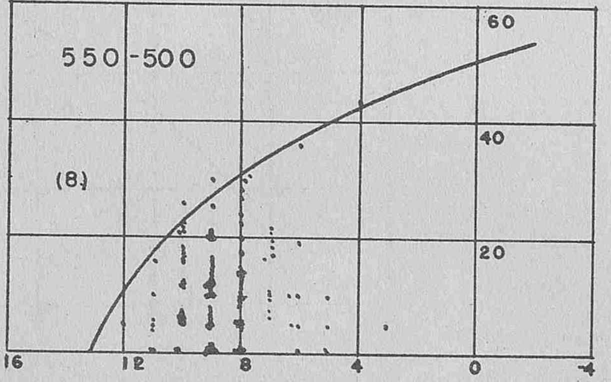
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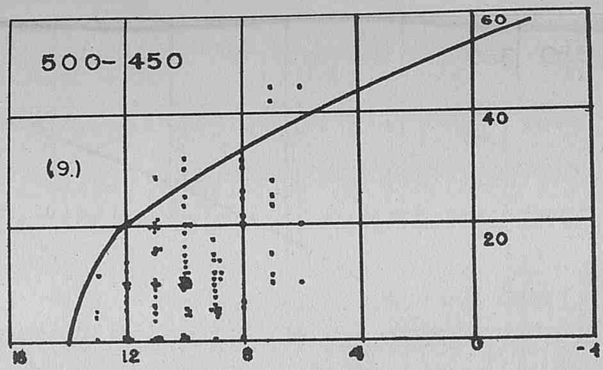
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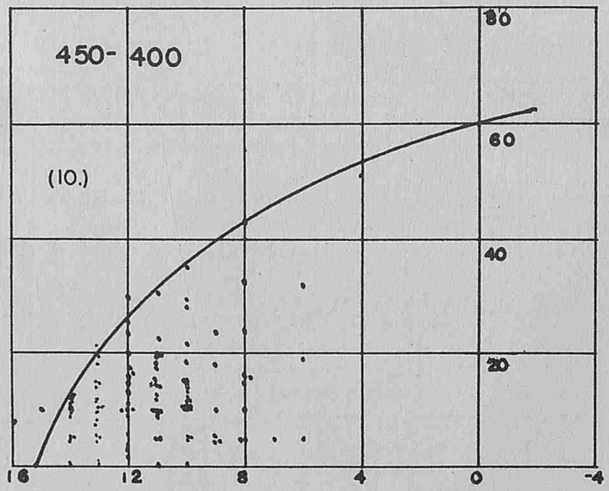
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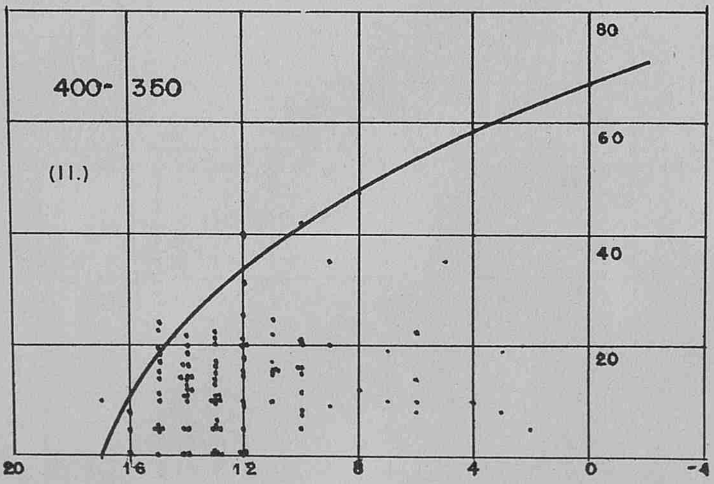
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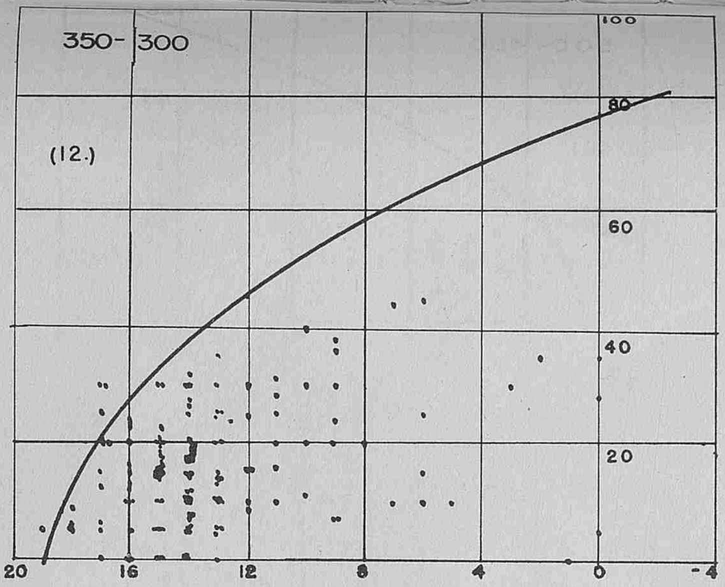
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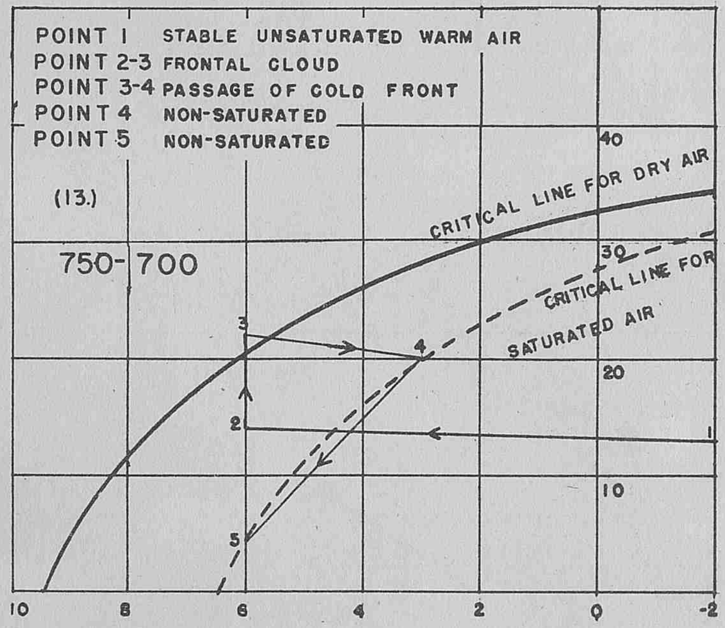
POINT 1 STABLE UNSATURATED WARM AIR
POINT 2-3 FRONTAL CLOUD
POINT 3-4 PASSAGE OF COLD FRONT
POINT 4 NON-SATURATED
POINT 5 NON-SATURATED
(13.)
750- 700
CRITICAL LINE FOR DRY AIR
CRITICAL LINE FOR SATURATED AIR
1
2
3
4
5
40
30
20
10
10
8
6
4
2
0
-2

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ON THE APPLICATION OF THE RICHARDSON CRITERION TO [LARGE SCALE TURBULENCE IN THE FREE ATMOSPHERE.] – Navaer title

by

Sverre Petterssen and W.C.Swinbank.

(June, 1943)

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1. Introduction.

A brief report* on an investigation into the relationship between the wind shear and the static stability of the air in the layer immediately below the tropopause was presented to the M.R.C. in October, 1942. In the course of the ensuing discussion, members of the M.R.C. suggested that the investigation should be extended to cover the entire troposphere. This has been done particularly with a view to finding empirical relationships between stability and wind shear and comparing these with the results obtained by applying the Richardson criterion to actual observations. A further demand for investigating this problem arose in connection with an investigation of the turbulence conditions that obtain at the tropopause.

In the present investigation use has been made of the reports from the British radio sonde stations for the period December 1942 to March 1943. The accuracy of the D/F method of determining the wind in the free atmosphere depends to a large extent upon the geometry of each case, and a tolerance figure is allotted to each observation assessing the accuracy. Only reports of the highest degree of accuracy(i.e. vector errors less than 5 km/hour) have been used in this investigation. In all, 1556 simultaneous observations of lapse rate and wind shear have been investigated.

The observations have been grouped according to levels, and the results are illustrated in 12 diagrams in which a lapse rate is the abscissa and wind shear the ordinate. The diagrams contain also the limiting curve between decrease and increase in turbulence as computed from Richardson's criterion. Since wind observations are made at intervals of 50 millibars, the diagrams represent the observed relationship between the mean wind shear and the mean lapse rate through layers of 50 mb. depth. The limiting curve in these diagrams would then represent the limit between the conditions that determine decrease and increase in the large scale turbulence of the atmosphere.

The diagrams show that, with very few exceptions, for any given degree of stability the observed wind shear is less than that required by Richardson's theory to cause the turbulence to increase.

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*MRP.51.

If the assumptions underlying the theory were correct and applicable to large scale turbulence, the diagrams would indicate that large scale turbulence almost invariably tends to decrease. Since this result appears improbable, it is shown that agreement between theory and observation can be obtained by dropping one of the assumptions made by Richardson; namely that the coefficient of eddy transfer of heat is equal to the coefficient of eddy transfer of momentum. The diagrams then lead to the conclusion that the former coefficient is smaller than the latter, the ratio between them being about 0.65.

2. Brief Summary of Richardson's Theory.

The role played by the eddy stresses that are brought into being when the motion of an incompressible fluid ceases to be laminar, and becomes turbulent, was first discussed by Reynolds [*], who assumed that the work done by the eddy stresses is consumed in two ways, viz.

(a) in increasing the kinetic energy of the turbulent motion,

and

(b) in increasing the intrinsic heat energy of the **fluid** by dissipation due to its molecular viscosity.

Reynolds showed that the kinetic energy of eddying will increase when the work done exceeds the heat energy thus created, and furthermore, that the rate of loss due to molecular viscosity becomes greater, and, therefore, the rate of gain in eddying energy less with increasing eddying velocity and decreasing size of eddies.

L. F. Richardson[†] extended the work of Reynolds to the case of a compressible fluid, and, in particular, to apply to the free atmosphere, where the eddies, in their vertical motion, must do work on account of expansion.

Richardson considers a large block of air, and, applying the principle of conservation of energy, determines the gain in eddying energy by estimating the losses from all the other available forms, namely:

(1) the intrinsic heat energy.

(2) the gravitational energy.

(3) the kinetic energy of the mean (i.e. non-turbulent) motion.

(4) the energy conducted to the block from the underlying surface.

(5) the energy due to radiation exchanges.

(6) the work done by pressure forces at the vertical boundary surfaces of the block.

By choosing the block sufficiently large, Richardson shows that (6) is negligible as against the energy derived from other sources, and by assuming that the temperature distribution is known, he is able to eliminate the effects of radiation. The sources (1) and (2) above can be treated as one source, which is hereafter referred to as heat energy. The loss of energy due to molecular viscosity is shown to be negligible except, perhaps, for very small eddies. Thus, if the diameter of the eddy were 10 metres, the loss of energy through this agency would amount to only 20 per cent in 24 hours.

Richardson next considers the mixing by the eddies of portions of air at different temperatures, and discusses two types of activity, which determine the supply and removal of energy to and from the eddies. One of these, which is called the boundary activity, is shown to determine the rate at which heat is supplied to, or removed from the underlying surface. The second, which is called the body activity, determines whether or not heat energy is being converted into eddying energy.

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[*] Phil. Trans. A. t clxxxvi p.123.

[†] Proc. Roy. Soc. A, Vol.97, 1920.

From the equations relating these two activities, Richardson derives an expression for the rate at which heat energy is transformed into eddying energy, and shows that it depends upon the upward gradient of entropy per mass of the air. Finally, he combines the energy supplied by this means with that supplied by the mean wind, and arrives at the following equation:

$$\frac{\partial E'}{\partial t} - \frac{2\mu_c \pi^2 E'}{\rho l^2} = \iiint \mu\left[\left(\frac{\partial \bar{V}_x}{\partial h}\right)^2 + \left(\frac{\partial \bar{V}_y}{\partial h}\right)^2\right] dh.dA$$
$$- \iiint \frac{cg}{\gamma_p}\frac{\partial \sigma}{\partial h}\, dh.dA$$

Here, the symbols have the following meanings:

E' = kinetic energy of eddying
μ_c = coefficient of molecular viscosity
ρ = density of the air
l = a linear measure of the size of the eddies
μ = coefficient of eddy viscosity
c = coefficient of eddy conductivity
h = height
$\bar{V}_x$ & $\bar{V}_y$ = mean horizontal wind velocity along perpendicular axes
A = cross section area of atmospheric block
g = acceleration of gravity
γ_p = specific heat of air at constant pressure
σ = entropy per mass of the air.

Following G.I. Taylor [x], Richardson now assumes that the coefficient of eddy viscosity (i.e. the coefficient of eddy diffusivity of momentum) is equal to the eddy conductivity of heat; in other words that

$$\mu = c$$

The small term due to molecular viscosity is neglected and equation (1) reduces to

$$\frac{\partial E'}{\partial t} = \iiint c\left\{\left[\left(\frac{\partial \bar{V}_x}{\partial h}\right)^2 + \left(\frac{\partial \bar{V}_y}{\partial h}\right)^2 - \frac{g}{\gamma_p}\frac{\partial \sigma}{\partial h}\right]\right\} dh.dA$$

This means that the sign of the rate of change of the kinetic energy of eddying (E') of the block depends upon the sign of the integrand

$$\left(\frac{\partial \bar{V}_x}{\partial h}\right)^2 + \left(\frac{\partial \bar{V}_y}{\partial h}\right)^2 - \frac{g}{\gamma_p}\frac{\partial \sigma}{\partial h}$$

---

[x] Phil. Trans. A, 1914.

The last term in this expression can be transformed so as to contain the vertical gradient of temperature, and becomes

$$\frac{g}{T}\left(\frac{\partial T}{\partial z}+\Gamma\right)$$

where T is temperature and Γ is the dry-adiabatic lapse rate. In this form, Richardson's criterion states that the kinetic energy of the eddying within a layer will increase or decrease according to whether

$$\left(\frac{\partial \bar{v}_x}{\partial h}\right)^2+\left(\frac{\partial \bar{v}_y}{\partial h}\right)^2 \gtrless \frac{g}{T}\left(\frac{\partial T}{\partial z}+\Gamma\right).$$

The assumptions underlying this criterion refer partly to the nature of the energy transformations and partly to the numerical relation between the coefficients for eddy transfer of momentum and heat. While the assumptions concerning the energy transformations seem entirely justifiable, there is very little evidence in support of the view that the coefficient of eddy transfer of momentum is equal to the coefficient for eddy transfer of heat. If the one differs from the other, the right hand side of the above inequality would contain a factor differing from unity.

Equating the shear term to the stability term in the above inequality, we obtain the equation for the critical curve separating cases of increase from cases of decrease in turbulence. This curve has been drawn in the diagrams that will now be discussed.

3. Construction of Frequency Diagrams.

In the diagrams appended the abscissa indicates the decrease in temperature (°F) from the base to the top of a layer of air of 50 mb. Thus, the abscissa indicates lapse rate of temperature with regard to pressure, zero indicating isothermal conditions, negative values temperature increase, and positive values temperature decrease along the vertical.

The ordinate indicates the magnitude of the vector difference between the wind at the base and at the top of the layer, expressed in km/hour.

The parabolic curve in each diagram indicates the Richardson criterion as a function of lapse rate and wind shear. The curve intersects the abscissa in the point corresponding to the dry-adiabatic lapse rate. Since the Richardson criterion is expressed in terms of shear and lapse rate with regard to height whereas the observations give similar quantities with regard to pressure, the critical curve has been computed with the conversion from height to pressure that holds for the standard international (I.C.A.N.) atmosphere. The errors due to this standard conversion are negligible in comparison to the observational errors.

In view of the fact that it is difficult to assess the accuracy of the wind observations when the balloon is near the earth's surface the lowest layer included in the investigation is the one from 900 to 850 mb. The atmosphere from 900 to 300 mb. has been divided into layers of 50 mb., and a diagram made for each layer. On account of the frequent occurrence of the tropopause above the 300 mb. surface no diagrams have been compiled for the layers above this level. Each point in the diagrams represents a simultaneous observation of wind shear and lapse rate. A point in the diagram below the critical curve indicates that, according to the Richardson criterion, the turbulence tends to decrease, and a point above the curve indicates that it tends to increase.

The diagrams 1 - 12 show the frequency distribution of the points in relation to the critical curve, and diagram 13 refers to the behaviour of an individual point.

4. Discussion of the Frequency Diagrams.

The points in the frequency diagrams may be referred to three categories

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according to whether they are below, on, or above the critical curve. The number of points belonging to each of these groups is given in Table I.

TABLE I.

Pressure Intervals (mb.)	Number of points		
	Below	On	Above
900 - 850	103	2	12
850 - 800	112	1	4
800 - 750	131	0	6
750 - 700	139	0	1
700 - 650	143	0	6
650 - 600	135	0	3
600 - 550	142	0	3
550 - 500	133	0	3
500 - 450	105	0	7
450 - 400	104	1	4
400 - 350	100	0	6
350 - 300	145	0	5
SUM	1,492	4	60

It will be seen that less than 4 per cent of the points occur above the curve, indicating that in less than 4 per cent of the cases were the conditions such that the turbulence would increase; while it would tend to decrease in about 96 per cent of all recorded cases.

Since the observations cover a period of four months, it is reasonable to assume that the mean increase in turbulence over the period investigated is nil. Therefore, if the time-trends of the variation in turbulence were symmetrical, or such that, on the average, the turbulence increased over as long time intervals as it decreased, an equal number of increases and decreases ought to be recorded, with the result that an equal number of points ought to be found above and below the critical curve.

On the other hand, the possibility exists that the increase in turbulence sets in suddenly while the stabilising processes act slowly. If this were so, the chance for recording a case of increase would be less than that for recording a decrease, and the ratio of the number of recorded decreases to recorded increases would increase with the assymmetry of the process. The results shown in Table I could then be explained by assuming that the assymmetry is about 24 to 1.

Assuming a pronounced assymmetry in the process, it is evident that the intensity of the increase in turbulence must be larger than the intensity of the decrease, since the net change over an extended interval must be nil. The diagrams show, however, that the points suggesting an increase are very close to the critical curve while the centre of gravity of the points suggesting a decrease is far removed from that curve. This, again, would indicate that, on the average, the intensity of the increase is less than the intensity of the decrease in turbulence. The diagrams 1 - 12 are, therefore, consistent neither with the assumption of symmetry nor with that of assymmetry.

The ideal solution would be to study the variation in turbulence of individual volumes of air. Only in very few cases were the wind direction and speed such that the air could be traced from one station to another. The results of the investigation of those cases, which will be described later, were not conclusive. On the other hand, analyses of balloon oscillations have shown rather conclusively that the periods of increasing and decreasing oscillations are of about the same length and there is no evidence to indicate any appreciable assymmetry in the processes. It seems therefore that the paradoxical results shown in the diagrams 1 - 12 must be due to either or all of the following factors.

(a) the application of the theory to turbulence on such a large scale that layers of 50 millibars are affected may not be permissible.

(b) condensation processes may contribute to the turbulent energy and these and have not been accounted for in the theory.

(c) the assumptions underlying the theory might lead to numerical disproportions affecting the position of the critical curve.

5. The Scale Effect.

It is of interest to note that one of the assumptions upon which the theory is based is that the volume considered is sufficiently large for the work transmitted across its vertical walls to be negligible in comparison to the other forms of energy involved. This assumption, which seems highly plausible, does not indicate that appreciable discrepancies between theory and observation should be due to considering layers of air of 50 mb. depth.

If the depth of the layer considered were an important factor one would expect to find in the diagrams 1 - 12 a distortion increasing with decreasing pressure, since the thickness of the considered layers increases with decreasing pressure. The layer from 900 to 600 mb. is only about half as deep as is the layer from 600 to 300 mb. Nevertheless, it will be seen from Table I that those two layers contain about the same number of points above the critical curve. It appears therefore that the thickness of the layer does not seriously influence the results.

Strictly speaking, the application of the criterion to layers of 50 mb. depth is permissible only when both lapse rate and wind shear are uniform within each layer. Observations with multiple theodolites made at Larkhill show that, for the depths considered, the procedure is reasonable as far as wind shear is concerned, and examination of the Tephigrams indicates that in the large majority of cases, the same is true for the lapse rate. This is particularly true of the upper half of the troposphere.

6. Influence of Condensation.

If the air were saturated, the saturated-adiabatic rather than the dry-adiabatic lapse rate would enter as a parameter in the criterion for turbulence. In diagram 13, the upper curve is the critical curve for non-saturated conditions, and the lower one is the critical curve drawn on the assumption that the saturated adiabatic can replace the dry-adiabatic lapse rate in the Richardson criterion when the air is saturated. This simple substitution of the saturated adiabatic may not be strictly correct, but it will suffice for the qualitative discussion that follows.

The points in diagram 13 represent consecutive observations at 6-hour intervals of the relation between lapse rate and shear during a frontal passage

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accompanied by condensation. It will be seen that when the saturated adiabatic replaces the dry-adiabatic, the critical curve is lowered and the points can more readily appear on the side of the curve where increase in turbulence is possible.

If the critical curve corresponding to saturation were drawn in the diagrams 1 - 12, about one-third of all points would be found above the curve and below the one that corresponds to non-saturated conditions.

The frequencies of relative humidity in excess of 97 per cent at Larkhill during the period under investigation are:

900 mb.	17.7 per cent.
750 mb.	8.1 per cent.
500 mb.	3.4 per cent.

Taking the average frequency of saturated cases to be 15 per cent (as a high value) it will be seen that less than 5 per cent of all the points in the diagrams 1 - 12 would be affected by condensation and there would therefore, all in all, be a large surplus of points indicating decrease in turbulence.

The importance of saturation can be tested also in the following manner:-

At the low temperatures occurring in the uppermost layers (say above 400 mb.), the saturated adiabatic is almost identical with the dry-adiabatic, and only minute amounts of energy are provided through condensation processes. The critical curve corresponding to saturated conditions would not differ noticeably from that valid for non-saturated air. The large deficit of points above the critical line at these levels can, therefore, not be ascribed to the influence on the position of the critical curve caused by condensation.

7. The Coefficients of Eddy Transfer.

The critical curves in the diagram were drawn on the assumption that the coefficient of eddy transfer of heat is equal to that of eddy transfer of momentum. If the former were smaller than the latter, the critical curve would be lower, and the number of points above the critical curve would be larger than shown in the diagrams 1 - 12. Since there is very little evidence in support of the view that the ratio of the said coefficients is unity, we shall now endeavour to determine the ratio by assuming that, during the period under investigation, the net gain in turbulence was nil. If this is so, then the true critical curve must be represented by the curve through the mean ordinates of the points. By drawing the smoothed curves for the mean ordinates in each of the diagrams 1 - 12 the values shown in Table II for the ratio of the two coefficients result.

It will be seen from Table II that there is no significant variation with elevation, the ratio being about 0.65 throughout. For turbulence on the scale here considered, it may be assumed that the ratio of the coefficients is about 2/3.

With this value for the ratio of the coefficients, a new critical curve may be drawn in all the diagrams. Relative to this new critical curve 42 per cent. of the points will be found to indicate a tendency for increase and 58 per cent of the points a tendency for decrease in turbulence. This assymmetry might suggest that on the average, the periods during which the turbulence increases are somewhat shorter than are those during which it decreases. With this assymmetry of about 4 to 6, the average intensity of the increase would stand as a ratio of the average intensity of the decrease as about 6 to 4.

TABLE II.

Ratio of the Coefficient of Eddy Transfer of Heat to the Coefficient of Eddy Transfer of Momentum.

Layer (millibar).	Ratio	Number of Observations
900 - 850	0.75	117
850 - 800	0.65	117
800 - 750	0.64	137
750 - 700	0.65	140
700 - 650	0.64	149
650 - 600	0.62	138
600 - 550	0.62	145
550 - 500	0.62	136
500 - 450	0.69	112
450 - 400	0.66	109
400 - 350	0.63	106
350 - 300	0.64	150
Mean for layer 900 - 550	0.653	1043
Mean for Layer 550 - 300	0.650	513

8. Time Variations.

An attempt has been made to study the variations in the turbulence of individual samples of air. Only in 44 cases were the wind direction and speed [illegible] such that the sample of air would move from one station to another during the [illegible] time interval that elapsed between two consecutive observations. The frequency distribution of these cases is shown in Table III, in which + stands for "increasing" and - for "decreasing", and zero means "no change".

TABLE III.

Lapse rate	Shear			Sum
0	-	0	+	
-	9	1	5	15
0	4	4	5	13
+	8	0	8	16
Sum	21	5	18	44

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It will be seen that a decrease in shear was as frequently associated with increasing as it was with decreasing lapse rate. The same appears to be true of increasing shear.

The apparent lack of correlation between the variations in wind shear and lapse rate seems to suggest that the time interval between the observations (i.e. 6 to 8 hours) is so large that the tendency at one instant is not appreciably determined by its value 6 to 8 hours earlier.

9. Conclusions.

The discussion in the foregoing paragraphs seem to indicate that relationships between wind shear and lapse rate in the free atmosphere are such that

(a) the Richardson criterion is applicable to large scale turbulence in the free stmosphere if the ratio of the eddy transfer of heat to that of eddy transfer of momentum is less than unity.

(b) the most probable value of this ratio is 0.65.

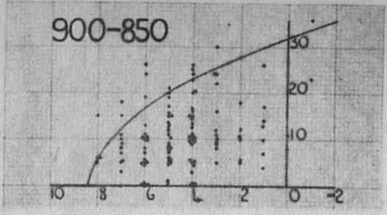
(c) above the friction layer this ratio is independent of elevation.

(d) condensation processes do not appear to have any appreciable direct effect upon the average conditions, although they may have an appreciable indirect effect through the lapse rates that they may create.

(e) the intervals through which the turbulence increases seems to be somewhat shorter than the interval through which it decreases, and that therefore the intensity of the increase is greater than the intensity of the decrease.

(f) the duration of the increase, or decrease, in turbulence in individual volumes of air is frequently less than 6 - 8 hours.

1.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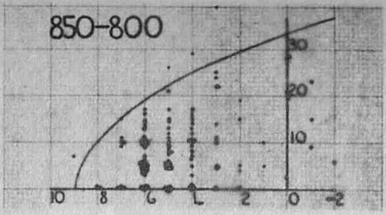
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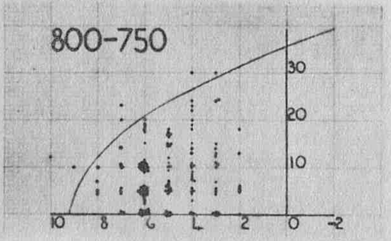
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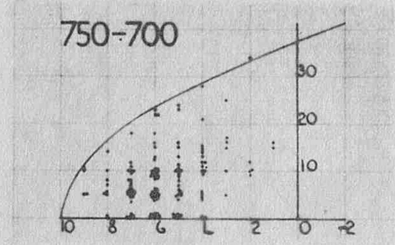
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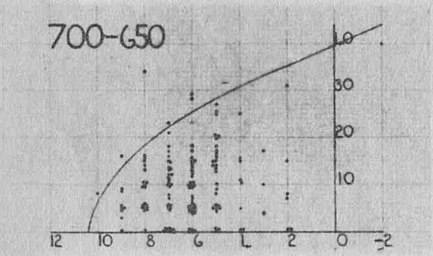
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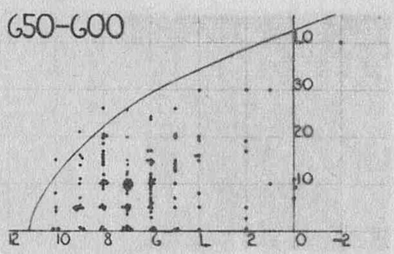
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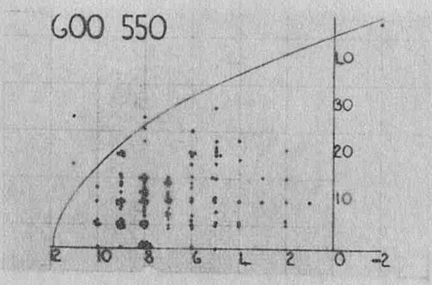
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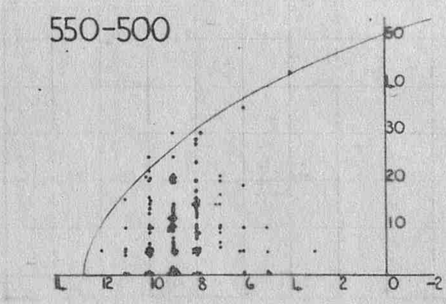
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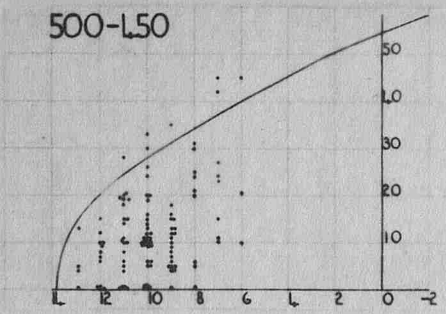
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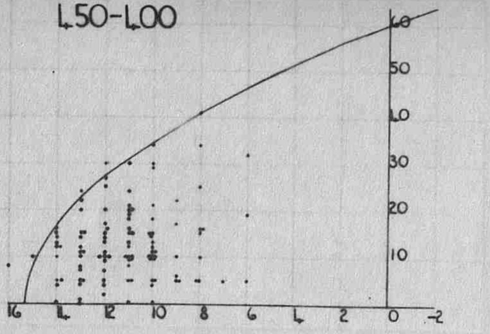
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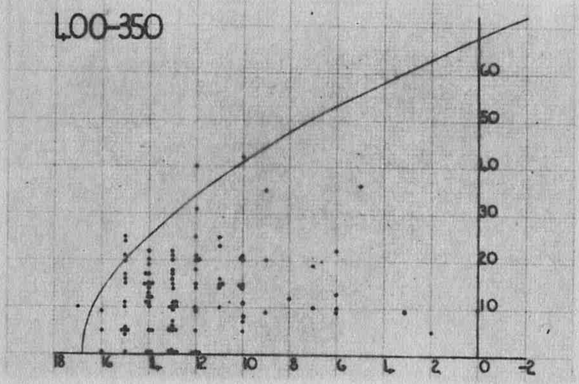
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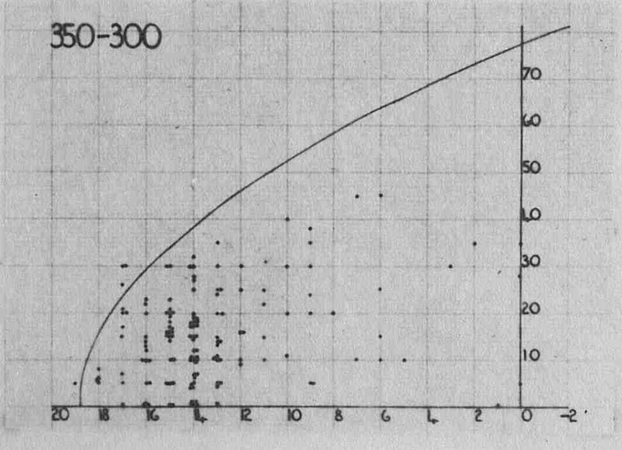
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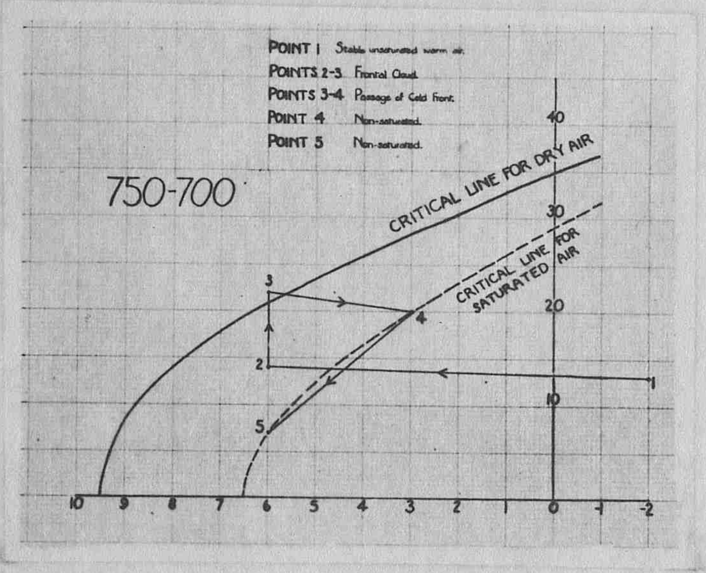
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POINTS 2-3 Frontal Cloud.
POINTS 3-4 Passage of Cold Front.
POINT 4 Non-saturated.
POINT 5 Non-saturated.
750-700
CRITICAL LINE FOR DRY AIR
CRITICAL LINE FOR SATURATED AIR
1
2
3
4
5
40
30
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9
8
7
6
5
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3
2
1
0
-1
-2

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AIR MINISTRY

M.R.P.130
16th August, 1943.

METEOROLOGICAL RESEARCH COMMITTEE.

Atmospherics in relation to Fronts and Air Masses.

Introductory Note.

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The observations of a single winter naturally do not provide a sufficient basis for the "Sferic climate" of winter. The conclusions in the paper must, therefore, be applied with caution even in the winter, and must not be taken to represent the conditions of other seasons.

Certain conclusions, e.g. the maxima in the frequency of atmospherics associated with fronts and polar maritime air masses in the central part of the North Atlantic are sufficiently consistent with what might be expected from the reasoning of meteorological physics, that they are believed to have a high probability of applying in similar conditions in any winter.

Similarly the broad conclusions as to the distribution of atmospherics in the neighbourhood of waves and fronts also have a high probability of being generally applicable to waves and fronts in the North Atlantic in winter.

An examination of the distribution of atmospherics in the summer months is being undertaken. As in these months diurnal convection will play a substantial part in the production of atmospherics, the geographical distribution and the distribution in relation to fronts and waves are likely to be materially different from those for the winter months shown in the charts and diagrams accompanying this paper.

M.R.P.130
16th August, 1943.

AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE.

ATMOSPHERICS IN RELATION TO FRONTS AND AIR MASSES.

(WINTER 1942-43)

by

F. A. Berson and Sverre Petterssen.

(June 1943)

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II. ACCURACY OF OBSERVATIONS

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I. INTRODUCTION.

Atmospherics is the word given to disturbances in wireless reception due to electric discharges in the atmosphere.

The place where the discharges occur is determined by the D/F instruments and technique developed by Sir Robert Watson-Watt and his collaborators*.

Simultaneous bearings are taken by the three stations: Dunstable, Leuchars and St. Eval, the lengths of the base lines being:

Dunstable - Leuchars	370 miles
Dunstable - St. Eval	220 miles

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* R.A.Watson-Watt, J.F. Herd and L.H. Bainbridge-Bell: Cathode Ray Oscillograph in Radio Research, Part 4, H.M. Stationery Office.

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St. Eval - Leuchars 425 miles

These base lines are sufficiently long to render reasonably accurate observations within the area indicated by the grid in Fig.1.

One of the principal purposes for which observations of atmospherics are made is to furnish information that will aid the meteorologists in the forecasting of thunderstorms and other instability phenomena.

The purpose of this investigation is to correlate the occurrence and distribution of atmospherics with the entities that enter into the synoptic analyses, such as fronts, air masses, cyclones etc.

The present investigation covers only the period from December 1st 1942 to March 10th 1943. During this period observations of atmospherics were made in runs as follows:

December 1st to February 14th.

1020 - 1040 clock time

1305 - 1325 clock time

1505 - 1525 clock time

February 15th to March 10th.

0905 - 0920 clock time

1105 - 1120 clock time

1305 - 1320 clock time

1505 - 1520 clock time

1705 - 1720 clock time

Thus, all observations were made during daylight hours, and the total of all runs investigated amounts to 107 hours.

Although the length of the series is much too short for a detailed study of the geographical distribution of atmospherics and their relation to fronts, air masses, etc., it has been found to be sufficiently long to form a basis for a study in broad outline of these features.

It should be noted that this report deals with winter conditions only; the results have, therefore, no bearing upon the type of atmospherics associated with heat convection over land in the warm season.

In winter, atmospherics are relatively rare over land (except along coast). Those discussed in this report are, therefore, mainly of oceanic origin. In the majority of cases, the atmospherics occur in distinct groups and travel with the fronts and air masses with which they are associated. On account of the few and short runs of observations and the lack of synoptic reports from the ocean areas, it has not been found profitable to study the behaviour of individual groups of atmospherics. The discussion below is therefore mainly statistical.

II. ACCURACY OF OBSERVATIONS.

The accuracy of the observations is not uniform within the area under investigation. The errors that make themselves felt may be referred to the following categories.

1. Errors in Reading. Particularly during periods of intense activity, it may be difficult to obtain accurate readings of the azimuth angles, since the flashes on the reading screen follow in rapid succession. Under normal conditions the reading errors average about ± 1°.

2. Site Errors. These are due to deviations of the radio waves caused by purely local conditions. The site error is usually constant for a given direction and wave-length, and although the site errors have not been determined for all directions, the inter-sections obtained on the atmospherics indicate that the residual errors are of the order of ± 1°.

3. Direction Finding Errors. These errors, which affect all direction finding and are believed mainly to be due to reflections from higher atmospheric layers, are most frequent at about sunrise and sunset; they are much less frequent by day than by night and their magnitude decreases with increasing wave-length. Since the runs of atmospheric observations were made in daylight and with a wave-length of 30,000 metres, the errors are limited to about 1°, except on rare occasions when it was obvious that reception conditions were poor.

4. Errors due to poor intersection. These errors tend to increase with the distance from the triangle of observing stations. The errors depend also upon the direction to the source and are largest when the source of discharge is so situated that the differences between the three azimuth readings are smallest. These errors are therefore largest over the western Mediterranean, the Iberian Peninsula and westwards towards the Azores; and they are almost entirely along the line joining the source to the centre of the triangle of observing stations.

III. GEOGRAPHICAL DISTRIBUTION.

The frequency of occurrence of individual atmospherics within squares of 10,000 sq. miles during the period of investigation is shown in Fig.1. The following features are apparent:

(1) The frequency is considerably higher at sea than over land (except in certain coastal regions), apparently due to the fact that active fronts and unstable air masses are more frequent over the oceans than over the continents in winter.

(2) On the North Atlantic Ocean a pronounced maximum occurs between about 45° and 50° North. The zone of maximum runs from the West-southwest to the East-northeast and is approximately parallel to the isotherms of the sea surface, the large majority of the atmospherics occurring on the warmer side of the 50°F isotherm.

(3) A secondary maximum occurs to the west of the Iberian Peninsula with a relative minimum along a line from Ireland to the Azores.

(4) In the Mediterranean area, the frequency is higher near the European coast than near the African coast.

(5) A somewhat distinct maximum occurs over England*.

(6) There appears to be a tendency for secondary maxima to occur along, or near, the windward coasts in the Prevailing Westerlies (e.g. Scotland, Norway and Iceland).

(7) Secondary maxima seem to occur in the vicinity of islands (e.g.

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* Most of these atmospherics occurred in connection with two fronts, and it is doubtful that a similar maximum would be maintained in long term means.

Shetland and the Faeroes). The secondary maximum in the vicinity of the Azores is less distinct. It should be noted, however, that the errors due to poor intersections are large in this area. Since these errors affect the recorded distance and not the direction to the source, it is plausible that the maximum near the Azores is more distinct than shown in Fig.1.

IV. RELATION TO FRONTS AND AIR MASSES.

A considerable number of the atmospherics recorded during the period under investigation occurred in connection with fronts, and the distribution of these has been correlated with young cyclone waves, occluded or partly occluded, cyclones and cold fronts.

1. Cyclone Waves.

Fig.2 shows diagramatically the position of the fronts in a young cyclone. The warm air (usually of subtropical origin) is to the south of the front and the cold air to the north of it. Most of the waves investigated occurred over the northeastern part of the North Atlantic near the normal position of the polar front.

In order to obtain a picture resembling the normal conditions, a common "system of co-ordinates" was imposed upon all waves, the one variable being distance from the apex of the warm sector expressed in miles and indicated by the circles, and the other variable being distance from the waving front, also expressed in miles.

The figures in the various areas of Fig.2 represent the number of atmospherics recorded.

It will be seen that a maximum of atmospherics occurs near the apex of the warm sector and that more atmospherics occur along the warm front than along the cold front. Almost no atmospherics were recorded along the more distant portions of the fronts. Apart from the atmospherics that occur in the warm air very near (in most cases, almost on) the warm front, none were recorded in the warm sector.

The absence of atmospherics in the warm sector proper seems to indicate that the tropical air arriving in the northeastern part of the North Atlantic in winter has acquired a considerable degree of stability.

In view of the fact that the frontal surface is a layer of pronouncedly stable air, it is evident that the atmospherics recorded near the fronts on their cold sides cannot be due to convection starting from sea level, since such convection could not penetrate the frontal surface and, therefore, not reach up to such heights as are recognised to be necessary for electric discharges to be released. The atmospherics recorded near the fronts must therefore be almost entirely due to latent instability (also called convective instability).

This type of instability is due to the distribution along the vertical of moisture, and is characterized by the potential wetbulb temperature decreasing with elevation. Active instability and convection are, under these circumstances, released when the air has ascended to such an extent that a layer of air has become saturated and part of the moisture condensed.

The distribution of atmospherics shown in Fig.2, therefore, seems to indicate that, on the average, the ascentional velocity of the warm air in nascent cyclones is most intense at the very apex of the warm sector, and furthermore, that it is more intense along the central portion of the warm front than along the central portion of the cold front. It also appears that atmospherics occur very rarely along the fronts at distances exceeding 400-500 miles from the apex of the warm sector.

Fig.2 contains a number of atmospherics at a considerable distance from the front at sea level. At least some of these may have developed within the cold air. With a slope of the frontal surface of 1/100, at a distance of 400 miles from the front, the frontal surface would be about 20,000 ft. above sea level, and it would not hinder deep convection from developing in the cold air. As will be shown in the discussion of cold fronts, purely frontal atmospherics are unlikely to occur at distances from the front exceeding 200-300 miles.

The atmospherics that occur in connection with cyclone waves appear normally as rather distinct concentrated and elongated groups, and travel with the speed of the wave. The marked tendency for the atmospherics to be concentrated at, or near, the apex of the warm sector is significant for the identification of young waves. Some of the most marked cases of such groups were recorded even before the wave developed closed isobars.

2. Occlusions.

A considerable number of the atmospherics recorded occurred in connection with occluded cyclones, and these are shown in Fig.3.

It appeared that the distribution of these atmospherics could more readily be co-ordinated with the "occlusion point" than with the pressure centre. In Fig.3, therefore, the fronts are drawn diagramatically as they normally occur in a partly occluded cyclone. The circles indicate distances from the occlusion point in steps of 100 miles. The set of curved lines indicates distances from the fronts, each zone being 100 miles wide. The numerals indicate the number of atmospherics observed within the significant areas, except that the numerals in the bottom row (i.e. below the heavy curve) are the sums of atmospherics recorded within each zone parallel to the fronts.

It will be seen that the atmospherics in typical occlusions occur much more frequently along the non-occluded than along the occluded part of the system. The highest frequency is found along and in the rear of the cold front to the south, or southwest, of the occlusion point. A considerable frequency is found in the apex of the warm sector, suggesting convergence and ascentional motion of the warm air.

The sums of frequencies in the bottom row show a marked maximum immediately in the rear of the cold front, a minimum about 200 miles in the rear of the cold front, with a secondary maximum about 500 miles within the cold air. A similar secondary maximum occurs also in connection with typical cold fronts and the origin of these atmospherics will be discussed in the following paragraph. It suffices here to remark that the atmospherics which appear in Fig.3 at distances from the cold or occluded front exceeding 200 to 300 miles develop within the polar air and are not directly associated with the fronts.

3. Cold Fronts.

The distribution of atmospherics in the vicinity of cold fronts is shown in Fig.4. The co-ordinate lines in this diagram are the distance from the front (in miles) and degrees of latitude. Thus, each quadrilateral is 100 miles wide and 2½ degrees latitude tall. The numerals within the quadrilaterals are percentage frequencies of atmospherics. The sums of such frequencies within zones of latitude are indicated in the right hand column (i.e. to the right of the heavy line), and the sums within zones parallel to the front are in the bottom row (i.e. below the heavy line). The cold fronts investigated are such as may be characterized as principal cold fronts, i.e. with a major outbreak of polar air in their rear, the cold air usually originating from the North American continent or the Greenland region. The wind in the rear of these fronts had always a westerly component.

It will be seen that about two-thirds of all the atmospherics that developed in connection with, or in the rear of, principal cold fronts occurred in the zone between 42½ and 50°N. It appears, therefore, that the winter air originating from the Arctic and American polar continental sources does not normally acquire the thermal structure (i.e. steep lapse rate, high temperature and high humidity) that is required to produce atmospherics until it has passed at least the 55° latitude circle. On its further journey toward the south and east the polar air attains a maximum of instability at about 45°N, and, as it streams farther to the south, the instability decreases rapidly. This rapid decrease in the instability south of about 42°N is presumably due to subsidence that tends to develop in the regions normally occupied by the subtropical anticyclones.

The distribution of the frequency of atmospherics as a function of distance from the front is also interesting. As in the case of occlusions, we find a maximum of frequency at, or immediately in the rear of, the front, a minimum between 200 and 300 miles from the front, and a secondary maximum between 300 to 500 miles within the cold air.

In Fig.5 is shown the percentage frequency of atmospherics as a function of distance from cold fronts and cold front type occlusions. The primary maximum appears to be due to frontal activity, and the secondary maximum to convection within the cold air in the rear of the front. The broken lines have been drawn so as to indicate the most probable frequency curves of frontal and air mass atmospherics separately. On the basis of these curves, it is reasonable to assume that the purely frontal atmospherics do not normally extend further than about 200 miles in the rear of the cold fronts. At a distance of about 300 to 400 miles in the rear of the cold front, the polar air usually attains a maximum of vertical extent, allowing convection to reach up to considerable heights, thus giving rise to a secondary maximum of frequency of atmospherics at about these distances from the front. At still greater distances the frequency decreases rapidly, apparently on account of the anticyclonic conditions, subsidence, etc. that tend to develop at these distances.

4. Polar Maritime Air.

Taking a distance of 200 miles from the cold fronts as the most probable limit between cold front and polar air mass atmospherics, the latter type may be separated from the former. The geographical distribution of the atmospherics that occurred in outbreaks of polar air (from the Arctic or the American polar continental sources) is shown in Fig.6. The maximum frequency of these atmospherics occurs at a distance of about 1,000 miles from the coasts of the cold continents over which the air originated, thus indicating that a very long sea track is required to cause atmospherics to develop in this type of polar continental air in winter. The rapid decrease in frequency south of about 45°N is suggestive of development of subsidence in the polar outbreaks that reach these latitudes.

5. European Polar Air.

In the later part of the winter, polar continental air from Europe streamed out over the Western Mediterranean, the Iberian Peninsula and the adjacent part of the North Atlantic. The heating of this air as it arrived over warm water caused numerous atmospherics, the distribution of which is shown in Fig.7. A pronounced maximum occurs along the Mediterranean coasts of France and Spain, and another maximum is found about 100 to 200 miles off the coast of Portugal. The thunderstorms giving rise to this latter maximum occurred mostly with easterly winds, suggesting that, for this type of air, a sea track of only 100 to 200 miles is frequently sufficient to cause considerable instability. On the other hand, as was shown in the foregoing paragraph, a sea track of 500 to 1,000 miles is needed for the polar air from Greenland and Labrador to develop appreciable amounts of atmospherics. This difference in the length of the sea track is apparently due to the difference in temperature between the two types of air masses. The higher the temperature the larger is the difference between the dry and the saturated adiabatics, and the amount of energy released through condensation is proportional to this difference.

6. Maritime Tropical Air.

During the period under investigation only a few cases were recorded; these appeared in rather confused patterns, and they were far too few in number to be investigated statistically.

7. Frequency Distribution of Frontal Atmospherics.

The frequency distribution of frontal atmospherics is shown in Fig.8. A rather uniform zone of maximum extends from the west-southwest to the east-northeast toward the English Channel. This zone is rather to the south of the zone of maximum frequency of fronts*, particularly in the eastern part of the Atlantic. Since latent (or convective) instability in the warm air is essential for the development of frontal thunderstorms in winter, the frequency distribution seems to suggest that the latent instability normally perishes when tropical air moves north of about 45 to 50°N.

8. Periodicity.

Since the occurrence of atmospherics over the North Atlantic Ocean is closely related to the position of the polar front and outbreaks of polar or arctic air, and since it has been suggested that there is a tendency for the frontal activity and the polar outbreaks to occur in a rhythmic fashion with a mean period of about 5 days, it is of interest to investigate whether there are indications of a similar periodicity in the occurrence of the atmospherics.

Two areas, A and B, were investigated separately. The area A (45° to 55°N and 20° to 35°W) obtained its atmospherics from the polar front and outbreaks of polar air, whereas the area B (35° to 45°N and 5° to 20°W) was under the influence of European air streaming from an easterly direction. It is, therefore, to be expected that if there is a periodicity of about 5 days in the polar front activity, it should make its influence felt in the number of atmospherics within the area A but not in the area B.

A period of analysis of the occurrence of atmospherics showed that:

(a) there is a very marked tendency for a period of 5 days to persist in the number of atmospherics within the area A.

(b) there is no evidence of a five-day period within the area B.

(c) there is some slight evidence of a period of 10 - 12 days in both the areas A and B.

The series of observations is, however, too short for a reliable determination of the longer periods.

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* S. Petterssen: Fronts and Frontogenesis in the Atlantic Area. Mem. Met. Assoc. I.U.G.G. 1936.

AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE.

REPORT
ON AN EXPERIMENT WITH AVALUATION OF GEOSTROPHIC DEPARTURES BY THE AID OF OBSERVATIONS FROM TRIANGLES OF RADIO SONDE STATIONS.

INTRODUCTORY NOTE.

The following discussion is based on a rigid mathematical treatment of a specific problem. The conclusion drawn must, therefore, be interpreted with this limitation in mind.

The result of the examination clearly indicates the limitation of what can be obtained from a single triangle of stations when the observations are subject to the errors that at present exist in the simultaneous measurement of pressure and temperature.

The results of the examination show that large errors in the computed geostrophic wind will occur if the pressure force is determined solely by reference to the observations at three points, without smoothing on the principle of continuity in space and time. Errors similar in character, though smaller in degree, would be found if the geostrophic wind at sea level were determined from barometer readings at three points only.

In practice the application of the principle of continuity in space and time enables the analyst to effect a large diminution in the errors. Furthermore, just as the wind at the surface assists the analyst in drawing the isobaric charts for sea level, so the wind at higher levels assists in delineating the pressure distribution at these levels, and the smaller the number of stations from which observations are available the greater is the assistance to be derived from wind observations. However, this mutual adjustment of the wind and pressure observations leads to a smoothing through which the relatively small departures of the actual wind from the geostrophic wind are largely eliminated.

M.R.P. 131
16th August, 1943.

AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE

REPORT

ON AN EXPERIMENT WITH EVALUATION OF GEOSTROPHIC DEPARTURES BY THE AID OF OBSERVATIONS FROM TRIANGLES OF RADIOSONDE STATIONS.

by

H. Christensen and Sverre Petterssen.

(June 1943)

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I. INTRODUCTION.

In M.R.P.128,"Notes on Upper Air Analysis", Part I, (January 1943), attention was drawn to the possibility of obtaining the geostrophic departures by comparing the actually observed wind with the geostrophic wind as obtained from observations of pressure and temperature in triangles of radio-sonde stations. Although the odds against obtaining satisfactory results were thought to be quite high, it was felt that an experiment with such evaluation was justified, particularly since positive results might prove to be of very great importance for the dynamical interpretation of the upper air chart, and, on the other hand, negative results might throw light upon the general accuracy of the observations.

In the first place, some interesting situations were chosen for evaluation, and the departures of the actual wind from the geostrophic wind thus determined were frequently found to be unreasonable both as far as magnitude and direction were concerned. The same was found to be true even when the investigation was limited to wind reports of the highest accuracy class, that is, reports with tolerance figure A (i.e. vector error less than 5 km. per hour). Since tolerance figure A was found useless as a criterion for sufficient accuracy, the investigation was turned toward the aim of exploring in a superficial fashion the causes of the discrepancies found. For this purpose use was made of the simultaneous observations of wind, pressure and temperature from the stations, Downham Market, Liverpool and Larkhill for the period March 15th to April 30th.

The results of this investigation seem to suggest:

(a) That the average accuracy of the observations is not sufficient for evaluation of geostrophic departures by the triangulation method.

(b) that the largest discrepancies are due to errors in the computed geostrophic wind, and not so much due to errors in the observed wind.

(c) that the errors in the geostrophic wind, as obtained by triangulation increase in frequency and magnitude with increasing elevation.

(d) that the major source of the discrepancies is the inaccuracy of the radio-sonde pressure and temperature observations.

II. THE METHOD OF TRIANGULATION.

In the daily synoptic routine, the contours of the pressure surfaces are drawn so as to render the best possible fit with the observed heights as well as with the observed winds. This adjustment, which involves the assumption of geostrophic balance, eliminates the possibility of obtaining the geostrophic departures by comparing the observed wind with the pressure force.

By the new radio-sonde technique winds and pressure gradients are obtained independently, and if the observations were sufficiently accurate, the geostrophic departures could be obtained by computing the pressure force from observations made in triangles of radio-sonde stations, and comparing the force thus obtained with the actual wind.

In the present investigation, use has been made of observations of wind, pressure and temperature in a triangle of stations, the base lines of which are as follows:

Larkhill - Downham Market	235 km.
Larkhill - Liverpool	248 km.
Liverpool - Downham Market	220 km.

It will be seen that the triangle is almost equilateral.

A triangular graph was constructed in such a fashion that the pressure force in the horizontal could be obtained by entering the observed differences in heights of standard pressure surfaces along the significant base lines. From the pressure force thus determined, the vector geostrophic wind was obtained. The diagram was constructed on a large scale so as to eliminate errors in the evaluation. The accuracy of evaluations will, therefore, depend entirely upon the accuracy of the observations of pressure and temperature from which the heights of the standard pressure surfaces are obtained.

The geostrophic wind vector determined by this triangle represents the mean vector within the triangle, and may, therefore, be taken to refer to the centre of gravity of the triangle.

If, $\mathbb{V}_1$, $\mathbb{V}_2$ and $\mathbb{V}_3$ are the actual winds as observed at the corners of the triangle, the mean vector wind within the triangle would be approximately

$$\mathbb{V} = \frac{1}{3}\left(\mathbb{V}_1 + \mathbb{V}_2 + \mathbb{V}_3\right)$$

since the triangle is so small that a linear distribution can be assumed.

Furthermore, if $\mathbb{G}$ is the geostrophic wind vector

$$\Delta\mathbb{V} = \mathbb{V} - \mathbb{G}$$

denotes the geostrophic departure. The acceleration ($\dot{\mathbb{V}}$) of the actual wind is at right angles to and to the right of the geostrophic departure.

The relation between the vectors $\mathbb{V}$, $\mathbb{G}$, $\Delta\mathbb{V}$ and $\dot{\mathbb{V}}$ is shown diagramatically in Fig. 1.

Having thus evaluated the departures, the accelerations are readily obtained. The difficulty then represents itself that there is no obvious way in which the computed accelerations can be tested; the evaluated accelerations might be largely influenced by errors in the observations. Since there is no way of testing the individual accelerations, the results will be tabulated in such a manner that certain probability tests can be applied.

Since the use of vectors is convenient in the tabulation of results, the evaluated departures will be expressed in a different manner, viz., let G be the magnitude of the geostrophic wind, V the magnitude of the actual wind, and β the angle from the geostrophic wind to the actual wind (see Fig. 1), the geostrophic departure, as expressed above, is uniquely determined by

$$\nu = G - V \qquad \text{and the angle } \beta$$

As will be seen from Fig. 1, ν is always smaller than the magnitude of the geostrophic departure ΔV, except when $\beta = 0$ or $\beta = 180^\circ$.

In terms of ν and β the departures may be classified as shown in Table I.

TABLE I.

TYPES OF BALANCES.

Type	ν and β		Description
(1)	$\nu = 0$	$\beta = 0$	Geostrophic motion
(2)	$\nu > 0$	$\beta = 0$	Gradient wind with cyclonic curvature
(3)	$\nu < 0$	$\beta = 0$	Gradient wind with anticyclonic curvature
(4)	$\nu > 0$	$\beta > 0$	Normal case of accelerated motion
(5)	$\nu \leqq 0$	$\beta > 0$	Acceleration; ν and β must be very small and the state of short duration. (This case should, therefore, appear infrequently in the random observations).
(6)	$\nu < 0$	$\beta < 0$	Normal case of retarded motion
(7)	$\nu \geqq 0$	$\beta < 0$	Retardation, otherwise as case (5) above
(8)	$\beta > 90^\circ$ or $\beta < -90^\circ$		Impossible

III. FREQUENCY OF DEPARTURES.

The triangulation method was applied to each of the standard pressure levels 900, 800, 700 etc. - 200 millibar. To abbreviate, the results have been summarized in Tables II and III for the lower and upper trosphere, respectively.

A mere inspection of these tables will convince the reader that appreciable errors are likely to occur quite frequently in the observations. Thus, for example, it is difficult to believe real geostrophic departures amounting to 100, or more, km. per hour can exist in the atmosphere. Particularly in Table III, such departures are quite frequent. It is equally difficult to have faith in the departures in direction (i.e. β) that exceed 90°. The great majority of the departures in Tables II and III are such that, if they were real, they would indicate acceleration. In view of the fact that the series of observations cover 47 days, it is difficult to believe that cases of accelerated motion should be encountered so much more frequently than cases of retarded motion. It appears, therefore, that there are both systematic as well as frequent accidental errors in the observations from which the departures were derived.

The departures shown in Table II and III have been referred to the 8 categories of balances shown in Table I, and the result of this re-grouping is shown in Table IV.

TABLE II.

Frequency of Departures v (km/hour) and β (degrees) in the 900, 800, 700 and 600 mb. surfaces.

v \ β	-100 to -180	-70 to -100	- 60	- 50	- 40	- 30	- 20	- 10	0	+ 10	+ 20	+ 30	+ 40	+ 50	+ 60	70 to 100	100 to 180	
-30 to -50											1	1					1	3
-20	1							1		1		1				1		4
-10		1				2	1	3	4	2			1			1		15
0	3	2			1	2	3	7	17	10	4	4	6	1		1		61
10						2	3	3	8	6	2	2	4	1	1	2		34
20	1							2	5	2	3	2	2			1		18
30									7	2	3	2	1				1	16
40								1	2	1		1		1			1	7
50								2	5	1	1		1					10
60							1		1		1			2		1		6
70 to 100									2	1			2	1				6
100 to 200											1	4						5
	5	3	0	0	1	6	8	19	51	25	16	17	17	6	1	7	3	185

TABLE III.

Frequency of Departures v (km/hour) and β (degrees) in the 500, 400, 300 and 200 mb. surfaces.

v \ β	-100 to -180	- 70 to -100	- 60	- 50	- 40	- 30	- 20	- 10	0	+ 10	+ 20	+ 30	+ 40	+ 50	+ 60	70 to 100	100 to 180	
Less than -50									2									2
-30 to -50	1	1							3	3	1	3		1				13
-20		1		1					2					1	1			6
-10		1			1			1			1			1				5
0					1	1		1	1				3	1			3	11
10		1						1		3				1		1	1	8
20			1			2	1	1	2	1	3			1				12
30					1	2			3	1	3	1		1	1	1		14
40						1	1		2	1	2	3	1	1			1	13
50	1										1	1		1				4
60							1		1			1	1					4
70 to 100									7	6	3	2	3	2			1	25
100 to 200	1					1		2	3	7	3	5	2	1			2	27
200 to 300								2	2	2	1	1	4			1	1	14
300 to 400										2		1						3
400												1			1	1		3
	3	4	2	1	3	7	3	8	28	26	18	19	14	12	3	4	9	164

It will be seen that a large percentage of the departures come in the category "impossible". If we to these add the departures in speed exceeding 100 km. per hour as highly improbable, we find that no less than 38 per cent of all observations above the 600 mb. level give departures in speed or direction, or both, that cannot be regarded as real.

The relatively high frequency of departures in the categories (5) and (7) must be regarded with suspicion, and, likewise, the high frequency of departures suggesting acceleration together with the low frequency of cases suggesting retardation seem to indicate that systematic errors have made their influence felt. The same is true of the frequencies indicating cyclonic and anticyclonic curvature.

The fact that positive departures in speed are so much more frequent than negative departures would suggest that either the observed winds are systematically, or very frequently, too low, or that the computed geostrophic winds are systematically, or very frequently, too high. The frequencies in Table III suggest that the errors are mainly in the computed geostrophic wind, since there is nothing to indicate that the actual winds could deviate from the observed wind by several hundred kilometres per hour. It seems, therefore, to be beyond doubt that at least the excessive departures in speed and direction are due to errors in the computed geostrophic wind.

TABLE IV.

PERCENTAGE FREQUENCY OF DEPARTURES IN THE VARIOUS CATEGORIES.

Category (according to Table I).	Percentage Frequency Lower Troposphere	Percentage Frequency Upper Troposphere
(1) Geostrophic balance	9	1
(2) Gradient wind with cyclonic curvature	16	12
(3) Gradient wind with anti-cyclonic curvature	2	4
(4) Normal case of accelerated motion	27	46
(5) Improbable or rare case of accelerated motion	17	10
(6) Normal case of retarded motion	4	2
(7) Improbable or rare case of retarded motion	15	12
(8) Impossible	10	16

Since the winds are predominantly from a westerly direction, the excessive positive departures in speed indicate that the evaluated meridianal pressure gradients are too high. Now, an error in the heights of the pressure surfaces at Downham Market would mainly influence the direction of the computed geostrophic wind, and not so much the speed. The high frequency of large positive departures in speed, therefore, suggest that either Larkhill reports systematically, or very frequently, pressures and temperatures that render too high values for the height of the pressure surfaces, or else Liverpool reports temperatures and pressures that give systematically, or very frequently, too low values for the heights of the pressure surfaces.

Finally, the fact that the frequencies of excessive positive departures in speed tends to increase with elevation, seems to indicate that the errors are such that they tend to accumulate with elevation. This tendency to accumulation is typical of errors in the temperatures.

IV. THE SOURCES OF ERRORS.

If V is the geostrophic wind, g the acceleration of gravity, λ the Coriolis parameter (i.e. $2\omega \sin \varphi$), and tan A is the slope of the pressure surface in the direction of the pressure gradient.

$$V = -g \frac{\tan A}{\lambda}$$

Let H_1 and H_2 be the heights of a pressure surface above two points on a line coinciding with the horizontal pressure gradient, and let B be the distance between these points, then

$$\tan A = \frac{H_2 - H_1}{B}$$

and substituting this in the above formula, we obtain:

$$V = -\frac{g}{\lambda} \frac{H_2 - H_1}{B}$$

In this formula, B is the length of the line within the triangle normal to the isobars, as shown in Fig.2. It will hereafter be referred to as the "effective base line".

Since in any given case, g and λ are constants, we obtain by varying the above formula:

$$\delta V = \frac{-g}{\lambda} \left[\frac{\delta H_2 - \delta H_1}{B} - \frac{H_2 - H_1}{B} \frac{\delta B}{B} \right]$$

$$\delta V = \frac{-g}{\lambda} \frac{\delta H_2 - \delta H_1}{B} - V \frac{\delta B}{B} \qquad (1)$$

Here, δV may be interpreted as the error in the computed geostrophic wind; it is composed of two parts, one that is due to erroneous heights of the pressure surfaces, and one that is due to errors in the length of the effective base line.

The base lines of the station triangle contain, of course, no errors. However, when the balloons ascend, the triangle representing the balloons will be deformed in proportion to the shear and convergence of the wind. Therefore the evaluated departures will be influenced by the height errors as well as by the base line errors.

1. The Base Line Error. It is easy to assess the nature of the base line errors.

Let H denote the height of the pressure surface in question, and a the ascentional velocity of the balloon. The time t required for the balloon to reach the height H is

$$t = H/a$$

We choose the x-axis along the contour of the pressure surface and the y-axis along the effective base line. Let u and v be the components of the horizontal velocity of the balloon (i.e. the wind velocity) along the co-ordinate axes. Then, during the time interval t the line joining two balloons released at a and a' in Fig. 2 will have increased from B to B_t such that

$$B_t^2 = B^2 + 2B\frac{\partial v}{\partial y}\frac{H}{a} + \left(\frac{\partial v}{\partial y}\frac{H}{a}\right)^2 + \left(\frac{\partial u}{\partial y}\frac{H}{a}\right)^2 \qquad (2)$$

Now, if the wind is geostrophic throughout the layer

$$v = \frac{\partial v}{\partial y} = 0$$

and the isobars would be parallel at all levels within the layer. The line joining the two balloons would have rotated an angle ψ such that

$$\tan\psi = \frac{\partial u}{\partial y}\frac{H}{a}\frac{1}{B}$$

and since the line joining the balloons would form exactly this angle with the direction of the pressure gradient, it follows that the effective base line suffers no change, except a translation, when the motion is geostrophic. The so-called base line error is therefore entirely due to the departure of the actual wind from the geostrophic wind.

In accelerated motion the effective base line suffers a translation, a rotation and a dilatation (or contraction). To the extent that the observed winds are reliable these changes can be evaluated. This has been done for a number of cases. Fig. 3 shows a case of quasi-geostrophic motion; the balloon triangle is displaced, but it is not appreciably deformed.

Fig. 4 shows a rather extreme case of horizontal wind variation. Liverpool (Lp) was situated in the centre of a depression, around which the winds were strong. It will be seen that the balloon triangle suffers a considerable rotation and a slight deformation during the ascent, but even in this rather extreme case the rotation and the deformation of the triangle are totally insufficient to account for such large departures of the actual wind from the computed geostrophic wind as were described in the foregoing paragraph.

It will be seen from eq. (2) that the base line error decreases with increasing ascentional velocity of the balloons, and the velocities used at the radio-sonde stations appear to be sufficiently large to reduce the base line errors to relatively insignificant amounts.

From the above discussion it appears that the major source of errors in the computed geostrophic winds are due to erroneous heights. Before proceeding to discuss the height errors some brief remarks will be made on the base line errors that occur when radio-sonde and aircraft ascents are used to form triangles.

The above discussion of the base line errors refers to triangles of radio-sonde stations only. If the triangle consists of aircraft ascents only, the base line errors vanish if the aircraft ascends vertically. However, if the triangle consists of radio-sonde and aircraft ascents, the base line errors may become very large. The horizontal motion of the balloon relative to the aircraft is then equal to the horizontal wind, and u and v may be substituted for $\frac{\partial u}{\partial x}$ and $\frac{\partial v}{\partial y}$ respectively, in equation (2). Since the balloons ascend with a speed which is about twice the ascentional velocity of the aircraft, a chaotic arrangement of the instruments will soon result if the winds are strong and the triangle small. To make an example, a mean velocity of about 100 km. per hour is frequently observed in the direction from Liverpool to Downham Market. By the time the aircraft at Downham Market has reached the 500 mb. surface, the balloon released from Liverpool will have reached well into the stratosphere and traversed about one half of the distance to Downham Market. Thus, two of the sides in the triangle of instruments could easily shrink, or stretch, by about 50 per cent during the ascents made from two radio-sonde stations and one aircraft station. Strict interpretation of such observations becomes very complicated when the triangles are small.

In this report only radio-sonde ascents have been used.

2. The Height Error. Since the base line errors have been found to be too small to account for the excessive departures that were discussed in Section 3, erroneous values for the heights of the pressure surfaces can be the only principal cause of the erratic geostrophic winds that resulted from strict triangulation.

According to Eq.(1) the error in the computed geostrophic wind due to erroneous heights may be written:

$$\delta V = -\frac{g}{\lambda}\;\frac{\delta H_2 - \delta H_1}{B}$$

The asymetry in the distribution of the values of the velocity departures in Tables II and III indicate that there is some cause of error which tends, in general, to give a systematic bias to $\delta H_2 - \delta H_1$. For an approximate analysis we assume the systematic part of the error to be constant (δH_c) throughout the series of observations. The resulting errors in the velocity departures may likewise be split into systematic and random components, by writing,

$$\delta V = \delta V_c + \delta V' = -\frac{g}{\lambda}\frac{\delta H_c}{B} - \frac{g}{\lambda}\;\frac{\delta H_2' - \delta H_1'}{B}$$

It is reasonable to assume that the random height errors (i.e. $\delta H_2'$ and $\delta H_1'$) are normally distributed about zero mean. If ' denotes the standard deviation

$$\sigma_V = \sigma_{V'} = \frac{g}{\lambda B}\sqrt{\sigma_{H_2}^2 + \sigma_{H_1}^2}$$

if H_2 and H_1 are of the same accuracy. A similar relation holds for the probable errors.

We may now introduce the "total range" instead of the standard deviation. Thus, if we exclude the 5% of the observations which represent extreme deviations (wild cases), the total range R is given by

$$R = 4\sigma$$

With $g = 10$, $\lambda = 10^{-4}$ and B = 200,000 m. (which is about the effective baseline that the triangle will give), we obtain for the total range of errors in V m/sec)

$$R = \frac{g}{\lambda B}\sqrt{2}R_H = 0{\cdot}7R_H$$

Where R_H = the total range in height error expressed in metres.

It will now be seen from Table II that a total range of 130 Km/hour which includes all but 8 out of 185 observations, could be attributed to a total range in height error of 50 metres. In other words, a random height error which was within 25 m. of the mean on all but 4 to 5% of occasions would account for the range of the velocity departures shown in Table II.

At the higher levels the exclusion of 8 observations would give a range of 350 Km/hr for speed errors, to which corresponds a range of about 150 metres in the height errors.

The mean values of the velocity departures in Tables II and III are about 20 and 80 Km/hr respectively. Putting these equal to δV_c , we find the corresponding values of δH_c to be 6 and 22 metres respectively.

The above analysis indicated the order of magnitude of the height errors required to account for the observed departures in speed. Accuracy beyond orders of magnitude cannot be obtained owing to the doubtful assumption of normal distribution of random errors and constancy of the systematic error. In this connection it may be noted that the observations from Liverpool are likely to be of lesser accuracy than those of the other stations, owing to site errors that have not been fully determined.

In order to evaluate the geostrophic departures by comparing the observed winds with the computed geostrophic wind it is necessary that the contours of the pressure surfaces be determined without smoothing by the aid of wind observations, for such smoothing tends to eliminate the departures to be determined. To effect a satisfactory smoothing of the pressure field without reference to the wind observations would require pressure and temperature observations from a dense network of stations, and even in such a network, the accuracy of the individual observation would have to be considerable.

Geostrophic departure less than 2 m/sec (7.2 Km/hour) are probably unimportant, and this value may, therefore, be taken as a definition of the accuracy that one should try to obtain. With the new radio wind measurement technique the actual winds will usually satisfy this requirement of accuracy. The question then arises: what accuracy is necessary in the pressure-temperature observations for the geostrophic wind to be determined with the required accuracy ? We will here aim at specifying accuracy limits within which the errors will lie on 19 out of 20 occasions. Using the term "allowable error" for these limits, the allowable errors of wind speed and of a single height determination will be $2\sigma_V$ and $2\sigma_H$, respectively.

Allowing an error of 2m/sec

$$2\sigma_V = 2$$

whence

$$\sigma_H = \frac{\lambda B}{9\sqrt{2}}$$

so that the allowable height error is related to the base line as follows:

Base line Km.	Allowable error in height (metres)
200	2.8
300	4.2
400	5.6
500	7.0
600	8.4

The accuracy of the pressure and temperature observation corresponding to these allowable height errors vary with elevation. Although errors may occur in pressure as well as in temperature they can for the purpose of height computation be referred to as errors in temperature only. The table below shows the accuracy with which the mean temperature (°F) of the air column must be determined in order to render heights of the accuracy indicated in the table above:

	Base line (Km.)				
Pressure surfaces. (mb.)	200	300	400	500	600
1000 - 750	0.6	0.9	1.2	1.5	1.8
1000 - 500	0.2	0.3	0.4	0.5	0.6
1000 - 300	0.1	0.1	0.2	0.2	0.3

/ The

The values given in this table may assist in defining immediate, or distant, aims of the radio sonde development, and indicate certain principles that should be considered in the building up of a suitable network of aerological stations.

V. THE VARIABILITY OF HEIGHTS.

In order to discover erroneous heights of the pressure surfaces, particularly in the upper atmosphere, graphs showing the heights of the 1000, 750, 500, 300 and 150 mb. surfaces as a function of time have been drawn, as part of the daily routine, since May 1943. A sudden variation in the height of a pressure surface over one station, which is not accompanied by similar, or consistent, variations at neighbouring stations is subjected to examination.

These height-time curves usually show a smooth variations at low levels, consistent with the development of the weather situation. With increasing elevation, the variations appear to become more and more erratic, and at the uppermost levels, the variation in height during six hour intervals are often larger than the variations from one day to the next, or the variations that seems to be associated with the development and travel of the pressure systems on the synoptic charts.

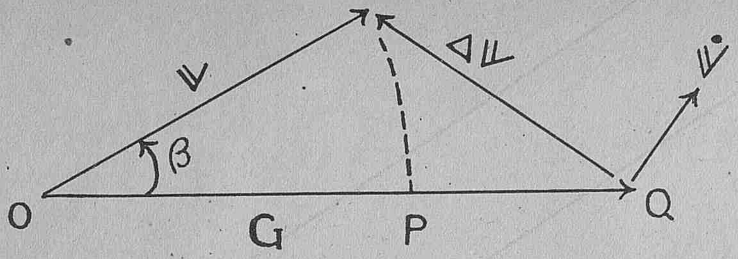
If these short period (six hour) variations were real, the interpretation would be that intense eddied travel through the upper atmosphere, and that their diameters are so small and their speed so high that they seldom make their influence felt at more than one station at a time, and rarely affect two consecutive observations at one individual station. The wind reports, however, do not substantiate the existence of such eddies, and it is plausible that these short period fluctuations are partly due to random errors in the observations.

Some of the statistical characteristics of these variations at Larkhill in June 1943 are shown in Table V. It will be seen that the ratio of the mean six hour variation to the mean 24 hour variation increases with elevation and becomes particularly large at the uppermost level.

TABLE V. Larkhill, June 1943.

Pressure levels (mb.)	1000	750	500	300	150
(1)Mean6-hour variation (ft.)	45	55	105	226	282
(2)Mean 24-hour variation (ft.)	91	94	170	344	365
(3)Ratio of (1) to (2)	0.50	0.59	0.62	0.66	0.77
(4) Permanency of sign in 6-hr. variations (%)	77	52	55	51	40

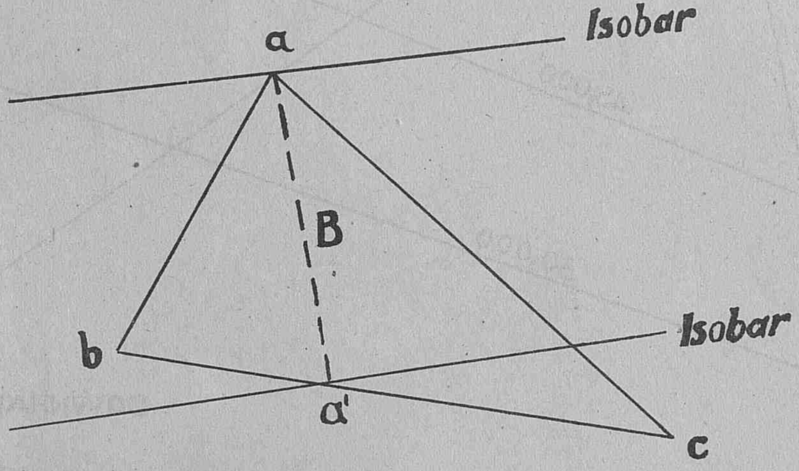
The permanency of sign of the variations shows some interesting features. At sea level (1000 mb.) the permanency of the sign in the six-hour variations is very high. At the higher levels, the permanencies resemble those characteristic of a chance distribution.



$\mathbb{V}$
$\Delta\mathbb{V}$
$\dot{\mathbb{V}}$
β
O
G
P
Q

$v = PQ$

Fig.1.



a
Isobar
B
b
Isobar
a'
c

Fig.2.

(F.2566). Wt.38704 550 11/43. Gp.959. FOSH & CROSS LTD.

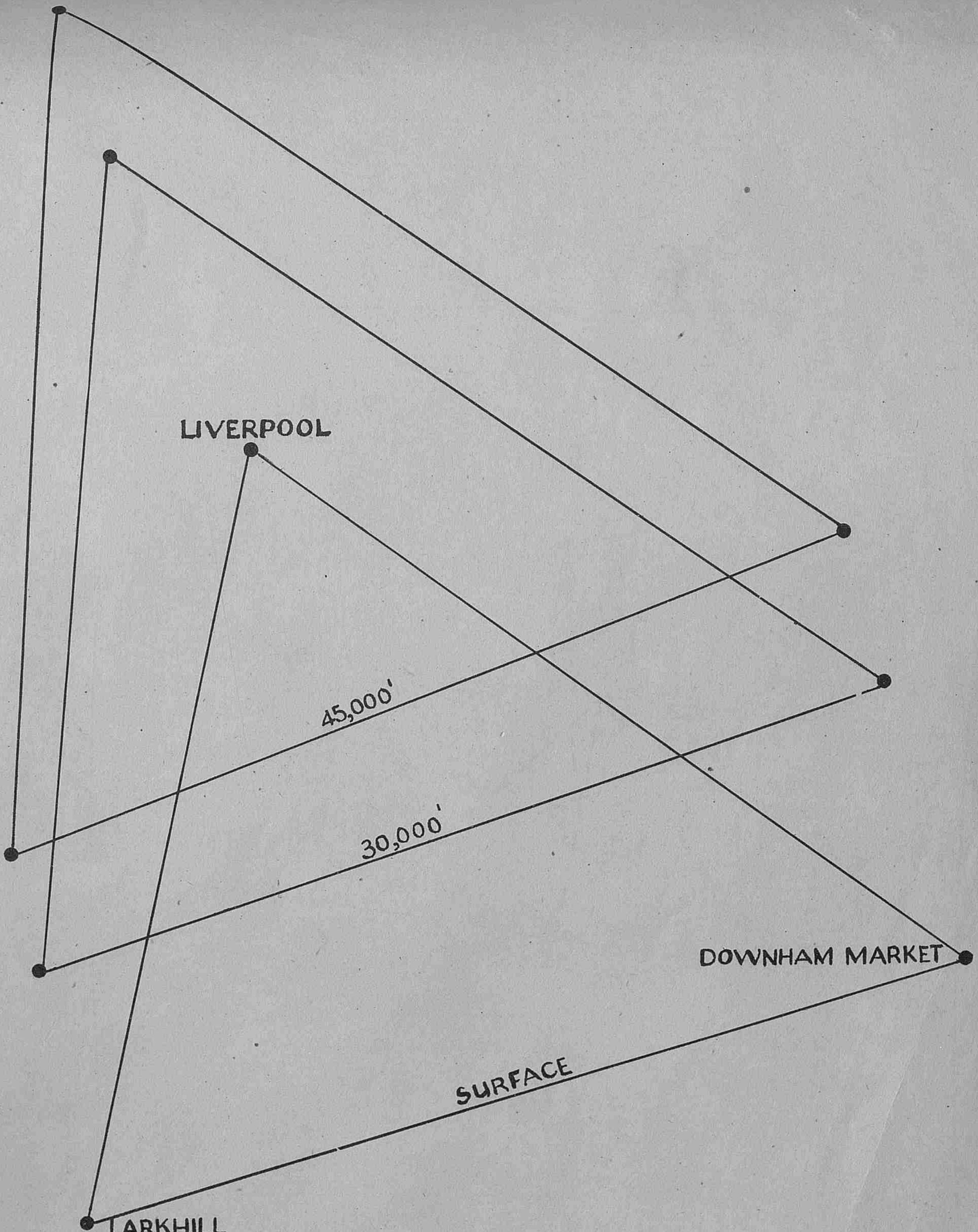

LIVERPOOL
45,000'
30,000'
DOWNHAM MARKET
SURFACE
LARKHILL

Fig. 3.

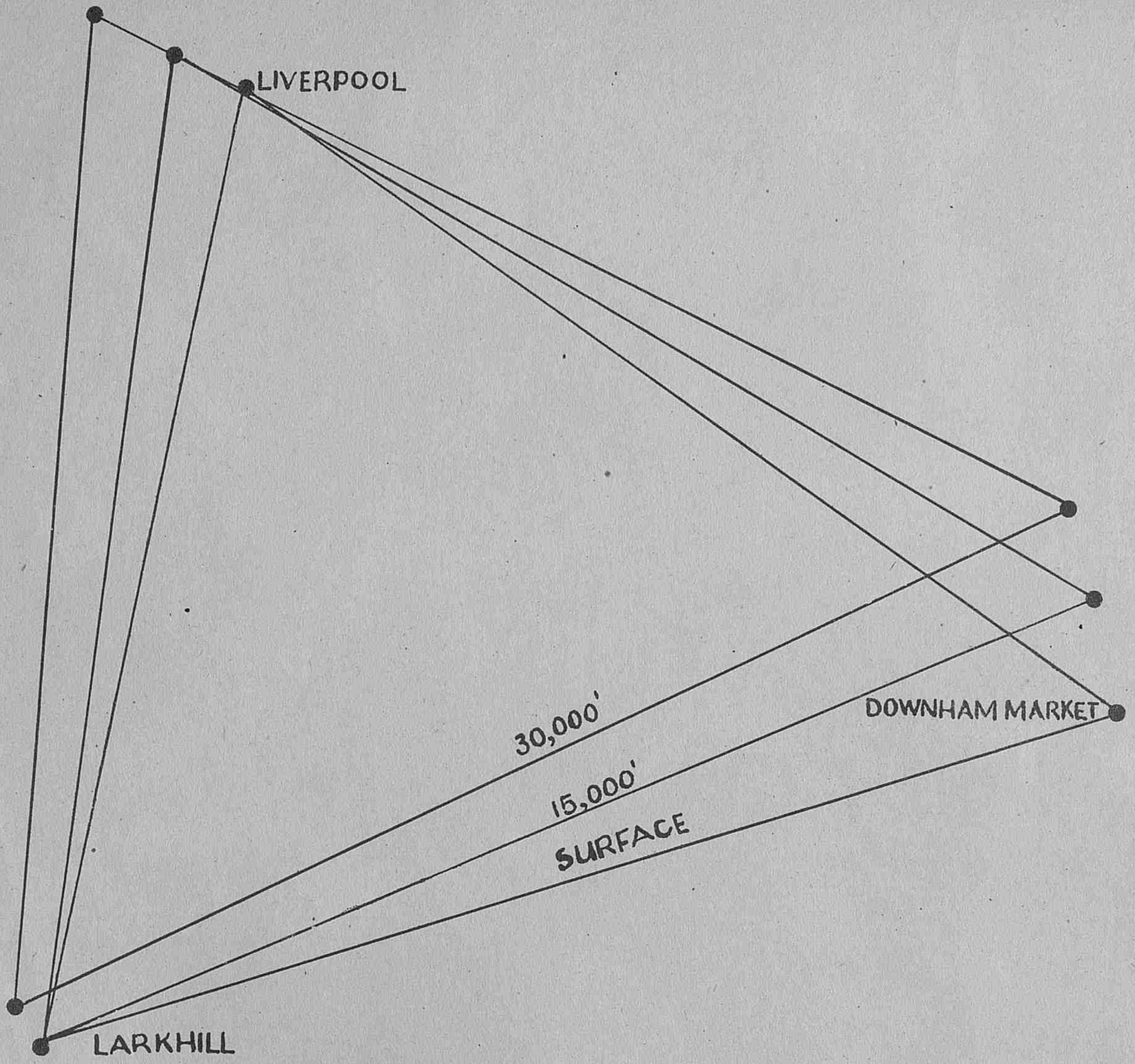

LIVERPOOL
DOWNHAM MARKET
30,000'
15,000'
SURFACE
LARKHILL

Fig. 4.

(F.2566). F. & C. LTD.

MRP. 141
16th September, 1943.

AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE

Variation of Wind with Height under Cyclonic and Anticyclonic Conditions.

by A.H.R. Goldie.

Some of the radio wind measurements made during the period October 1941 to March 1942 have recently been worked up by Miss F.M. Jones to obtain average wind speeds in winter at different heights under cyclonic and anticyclonic conditions respectively. The data used were for Larkhill, Fazakerley and Downham Market.

The mean wind speeds are shown in Fig.1. The means of $v\frac{\rho}{\rho_0}$, ρ being the air density at the level at which v is measured, and ρ_0 the density at sealevel are also shown; these latter curves show the relative air mass transport at different levels.

It is seen that in depressions there is approximate constancy of air mass transport (the "Egnell - Clayton law") in the region from 2 to about 8 km., suggesting complete vertical exchange of momentum in this region. Below 2 km. there is only partial mixing in the vertical direction and this is presumably in part connected with the restriction of movement arising from the fixed horizontal boundary (the ground), and in part with the convergence of air towards the centre of the depression, which takes place largely in the lowest 2 km.

Constancy of ρv with height involves constancy of barometric gradient with height, and this means that the horizontal barometric and thermal gradients bear the relation

$$\frac{\Delta\theta}{\theta} = \frac{\Delta p}{p}$$

i.e., level for level the air is colder towards the centre of the depression by an amount which considerably exceeds the adiabatic relation

$$\frac{\Delta\theta}{\theta} = \cdot 29\frac{\Delta p}{p}$$

This in turn points to greater ascent of air near the centre of the depression than in the outer region. The values of $\frac{\Delta\theta}{\Delta p}$ (a) as computed from adiabatic relations, (b) as derived from the statistical average relations (all cases) of W.H.Dines and (c) as estimated from the distribution of wind with height in cyclones in Fig.1 are as follows:-

	(a)	(b)	(c)
4 km.	.12	.46	.43
8 km.	.19	.47	.66

Above 8 km. and in the region of the tropopause there is also apparently not complete mixing with the levels below and the horizontal temperature gradients begin to reverse. The general distribution of velocities is in accord with a cell-like action in which the upward motion is mainly in the region from 2 to 8 km., whilst below 2 km. and again above 8 km. in the region of the tropopause the motion in the cell is only horizontal.

Under anticyclonic conditions the variation of speed with height is similar to that in depressions in the region from 5 to 8 km. and also above 8 km. It is however only from 4 or 5 to 8 km. that there is approximate constancy of air mass transport in anticyclones,

....

and thus a connection between the horizontal gradients of temperature and pressure of the form $\frac{\Delta\theta}{\theta} = \frac{\Delta p}{p}$ The weaker circulation in the anticyclone below 5 km. may be associated in part with the fact that in many anticyclones and wedges a discontinuity of air mass exists at about 3 km., the air below this level being relatively cold air, of northerly origin which is gradually leaking away from the system, whereas the air above this discontinuity may be warm air of sub-tropical origin. Since the level from 2 to 5 km. is sufficiently high to be practically unaffected by surface friction, it may be taken th there is in this level a definite change in barometric gradient, i.e., the gradient diminished from 5 down to 2 km. This means that, level for level, between 2 and 5 km., the air towards the centre of the anticyclone is very definitely warmer than that towards the outer fringes. It could be warmed purely as a result of greater subsidence near the centre, i.e., if it starts as a mound or heap of cold air it apparently flattens down most rapidly in the centre. But a further point is that subsidence of air through a given distance warms the air as a rule more than ascent through the same distance will cool air, since the former means rise of temperature according to the dry adiabatic and the latter in general involves fall of temperature only according to the moist adiabatic. It is chiefly this therefore that explains why in the anticyclone there is diminution of barometric gradient from 5 km. down to 2 km. (or lower) whilst this diminution is not shown in the cyclone.

Thus, from the wind distribution it can be deduced that isobaric surfaces are on the average also isothermals in the cyclone from 2 to 8 km. height, and in the anticyclone from 5 to 8 km. height; but in the centre of the anticyclone in the levels below 5 km. the isothermal surfaces should bulge upwards more than the isobaric surfaces and, somewhere under the 5 km. level there should be a closed isothermal, indicating an isothermal layer or discontinuity in the temperature lapse.

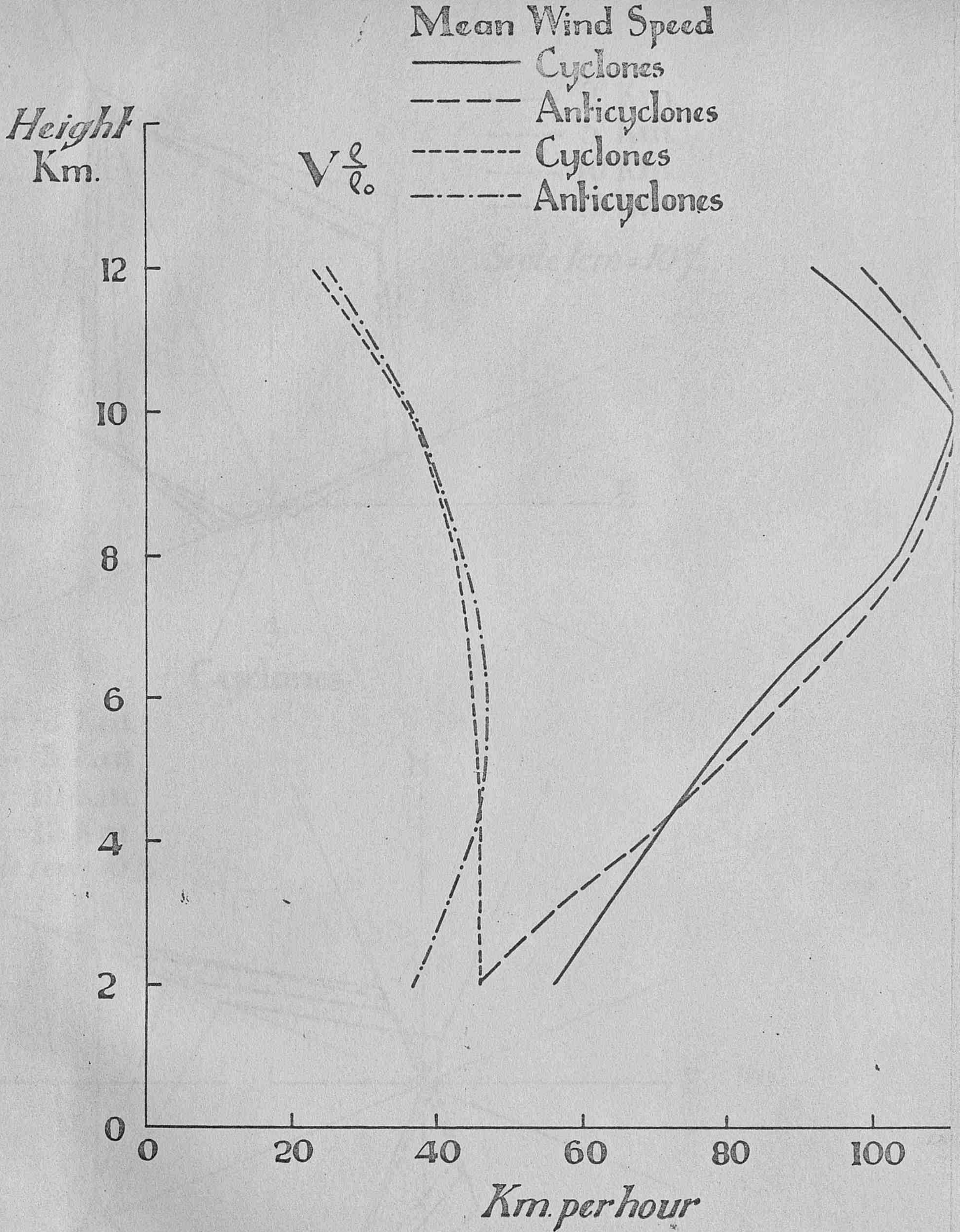
All the above remarks relate to average conditions, and the reasoning is based on the assumption of geostrophic balance.

The practical attack on dynamical problems, which is rendered difficult by the comparative roughness of upper air measurements, might be facilitated by the assumption of these results as average quasi-stable conditions, and by the investigation of departures from such conditions as being likely to be specially significant as regards development. The object would be to examine the vorticity of each depression as a unit. For example, increase of air mass transport with height in the levels from 2 to 8 km. except in so far as it was approximately balanced by extra coldness of air towards the low pressure would point to departure from geostrophic balance somewhere, either too low a speed in low levels in relation to existing barometric gradient, or too high a speed in high levels; in turn these conditions should involve increasing convergence towards the low pressure in low levels and divergence in high levels with ascent of air in the centre and increasing vorticity of the system until a condition of balance was attained. And correspondingly too cold air in the central area in relation to the variation of air mass transport with height would point to a system in which exhaustion of the kinetic energy was inevitable.

The approximate equality shown in Fig.1 in mean wind speed between cyclones and anticyclones above 5 km. is rather remarkable but is perhaps accidental. The averaged surface level pressures in the two groups of observations were 1007 and 1027 mb.

The distribution of wind directions is equally remarkable. The wind roses for Highs and Lows at 2, 5, 10 and 12 km. are given in Fig.2. The number of observations in each case at each level is around 270. This means that during the winter of 1941-42 practically all the Lows which could be specified were centered to the northward of the observing stations, and practically all the Highs were to the westward. This is another illustration of the difficulty of getting upper air information about all sides of either cyclones or anticyclones if only British observations are available.

FIG. I


Mean Wind Speed
——— Cyclones
— — — Anticyclones
$V\frac{\rho}{\rho_0}$
- - - - - Cyclones
—·—·— Anticyclones
Height Km.
0
2
4
6
8
10
12
0
20
40
60
80
100
Km. per hour

28 September 1943

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Gt. Brit. Met. research comm.

REPORT ON GOLD-BEATER SKIN HYGROMETER

by E. Gluckauf

A Practically all the fundamental data necessary for the construction of a GBS hygrometer are given in my 'Quarterly Report on work done for the Meteorological Office during July to September 1941'. The report gives figures for

(1) the calibration of GBS between +18° and -41°C

(2) the observed hysteresis effects after exposure to low humidities (observed at +12°, -21° and -28°C.)

(3) the lag of GBS between +18° and -68°C for air velocities varying from 1·2 to 6 m/sec.

(4) the influence of the relative humidity on the lag at constant temperature.

Some additional experiments which have been done later at an air velocity u of 34 m/sec. show the lag to be approximately inversely proportional to $\sqrt{u}$.

Taking all these data into account it follows for the lag (half way time τ mins.) at the absolute temperature T, the air speed u cm/sec. (under turbulent air flow), the average relative humidity with respect to water h (h=1 at 100 relative humidity) and a length of GBS of 10 cm:

$$\tau = \frac{10^{\frac{2000}{T}-7\cdot5}}{\sqrt{u}\,(h-h^2)} \qquad (1)$$

Very similar values to those obtained from this empirical equation can be deduced by purely theoretical calculations. A lower limit for the lag of GBS can be obtained from the knowledge of water content of GBS, its thickness and a number of known physical constants, under the assumption that diffusion within the skin can be neglected as compared with the diffusion resistance of the boundary layer.

If A = weight of 1 cm.^{2} of GBS 1·4 x 10^{-3} g/cm^2 at 50% RH

h = relative humidity as indicated by GBS.

$\frac{dA}{dh}$ = change of weight with rel. humidity = approx. 0·2A

= 2·5 x 10^{-4} at humidities of about 50 % RH.

D = diffusion constant of water valpour in air $= \frac{0\cdot25}{P} \times \left(\frac{T}{273}\right)^{1\cdot75}$ P in atm.

C_s = saturation concentration over water in g/cm^3

δ = thickness of boundary layer = $4\cdot5\sqrt{\frac{v.x}{u}}$

u = air speed in cm/sec.

x = distance from edge of GBS down wind.

v = kinematic viscosity

The change of weight at the time t after a sudden alteration of the humidity from h_0 to h_1 will then be

$$\frac{dA}{dt} = \frac{2 \cdot D \cdot C_s (h_1 - h)}{\delta}$$

and the change in indicated relative humidity will consequently be

over

$$\frac{dh}{dt} \qquad \frac{2 \, . \, D . \, C_s \, .(h_1 - h)}{0.2 \, . \, A . \, \delta}$$

from this follows for the time of the half change

$$\tau = \frac{0 \cdot 69 \, . \, A \, . \, \delta}{10 \, . \, D . \, C_s} \approx \frac{0.31 \, . \, A \, . \sqrt{\frac{v . x}{u}}}{D \, . \, C_s} \qquad (2)$$

taking the following numerical values, e.g.:

$x = 1 cm.$

$A = 1.35 \times 10^{-3}$ gr(drg)

$u = 3500$ cm/sec.

$T = 230^{\circ}K$

$C_s = 1 \cdot 5 \times 10^{-8}$ gr/cc

$v = 0 \cdot 10$

$D = 0 \cdot 18$

} at 230°K

we obtain a time for the half-change of 83 sec. The value experimentally determined was found to be 81 secs., while the interpolated formula (1) results in a value of 65 seconds.

Other measurements agree similarly well with the values* calculated from (2); for instance

T	u cm/sec.	(mins) observed	(mins) calculated
273	200	0.2	0.17
230	170	5	6·3
230	330	3·6	4·5
214	120	40	55
214	3400	7	10

Both from the experimental evidence and from theoretical considerations it would appear, therefore, that the GBS cannot be used at temperatures lower than -60°C., (even if the effect of the reduced barometric pressure at great heights is taken into account), if the time of measurement is limited to 2 - 3 minutes.

B The question, how far it is possible to extrapolate the equation (2)

$$\tau = \frac{0.31 \cdot A \cdot F}{D \cdot C_s} \sqrt{\frac{v \cdot x}{u}}$$

to air velocities and pressures beyond those experimentally tested, can be answered only within limitations. If one assumes that internal diffusion of water has the same coefficient for both hair and gold beater skin - their chemical composition is rather similar, except that they have a keratine and fibroine structure respectively - it becomes possible to make at least an estimate, which lag times form the lower limit imposed by internal diffusion.

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* Attention should be drawn to the fact that the actual measurements were made with GBS of 10 cm. length while the best agreement is reached if the lag is calculated for a length of 1 cm. Actually the two sets of figures are not strictly comparable as the calculations are based on a laminar flow of air, while in the experiments the air flow was kept specially in a highly turbulent condition. This difference in the state of the air flow apparently just compensates for the different lengths of the GBS.

Diffusion into a cylinder gives the half way position τ_c by equation

$$\frac{1}{2} = \sum_{0}^{\lambda=\infty} \frac{4}{n_\lambda^2} \cdot e^{-\frac{\alpha^2}{r^2} n_\lambda^2 \cdot \tau_c}$$

where α = internal diffusion coefficient
r = radius of cylinder
n = zero roots of Bessel function of zero order.

The numerical solution gives

$$\frac{\alpha^2}{r^2} \cdot \tau_c = 0.0664$$

Diffusion into a plane sheet of thickness l from both sides gives for the half way position

$$\frac{1}{2} = \frac{8}{\pi^2} \sum_{1}^{n=\infty} (2n-1)^{-2} \cdot e^{-\frac{\alpha^2 \pi^2 (2n-1)^2}{l^2} \tau_p}$$

which gives the numerical value

$$\frac{\alpha^2}{l^2} \cdot \tau_p = 0.0213$$

One must therefore expect that - given the constancy of in both materials -

$$\tau_p = 0.32 \cdot \frac{l^2}{r^2} \cdot \tau_c$$

The thickness of the GBS used is 1.3 x 10^{-3} cm, while the average radius of hair used for meteorographs is between 0.004 to 5, say 0.0045 cm. The limiting half way time of GBS should therefore be

$$\tau_{GBS} = 0.027 \, \tau_{hair}$$

Table 2 shows that the critical air speeds, at which the internal diffusion resistance becomes comparable with the diffusion resistance of the boundary layer, are at low temperatures and atmospheric pressure far too high for practical verification, while at the higher temperatures where internal diffusion comes into play at attainable air speeds the limiting lag times are so small that experimental checks are practically impossible.

Table 2

T	τ hair minutes found	τ GBS intern. calc.	τ GBS P=1, u=1.7 found	u crit. m/sec P= 1
293	0.7	0.02	0.063	17
273	1.4	0.04	0.2	40
253	3.3	0.09	0.8	130
233	9	0.24	3.6	400
213	(30)	0.8	25	1650
203	(57)	1.5	70	3700

On the other hand the reduction of air density at greater height will bring the internal diffusion into play at air velocities of 100 m.p.h. (45 m/sec).

over

Table 3 (all τ in minutes)

Height km	T	P atm.	lag intern.	at 100 miles/h boundary lag	at 100 miles/h total lag	at 200 miles/h boundary lag	at 200 miles/h total lag
0	276	1	0.04	0.03	0.07	0.02	0.06
3	258	.7	0.08	0.08	0.16	0.06	0.14
6	240	.47	.17	.19	.38	.13	.30
9	222	.30	.4	.6	1.0	.4	.8
12	218	.19	.6	.6	1.2	.4	1.0
15		.12		.4	1.0	.3	.9
12	213	.19	.8	1.0	1.8	.7	1.5
15		.12		.6	1.4	.4	1.2
12	203	.19	1.5	2.7	4.2	1.9	3.4
15		.12		1.6	3.1	1.1	2.6

It can be seen from table 3 that an increase of the air velocity above 100 miles/hour promises little improvement as regards lag, while the strain on the GBS will be increased quite out of proportion. It can also be seen that the limit of usefulness (half-way-change = 1 minute, 3/4-change = 2.5 minutes) of GBS can reach a temperature as low as 218°K, if the relative humidity is in the region of 50 RH which at this temperature is near ice saturation. At lower relative humidities, however, the internal lag will increase considerably, as it does in hair. It increases at 20 % RH and 10 % RH by a factor of 5 and 12 respectively, while the boundary lag increases only by a factor of 1.6 and 2.7 respectively. (see table 4)

Table 4
(u = 100 miles/h)

km	T	P	lag at 20 % RH int.	lag at 20 % RH bound.	lag at 20 % RH total	lag at 10 % RH int.	lag at 10 % RH bound.	lag at 10 % RH total
0	276	1	0.2	0.05	0.25	0.5	0.08	0.6
3	258	0.7	0.4	0.13	0.53	1.0	0.2	1.2
6	240	0.5	0.9	0.3	1.15	2.0	0.5	2.5
7.5	231	0.4	1.3	0.5	1.8	3.1	0.9	4.0
9	222	0.3	2.0	1.0	3.0			

Table 4 shows that at low humidities the usefulness of GBS (thickness 1.3 mg/cm^2, commercially called double thickness) becomes very much restricted. Though this thickness has up to now always been used for meteorological instruments for reasons of greater strength, it appears to be necessary to use, at low temperatures single thickness GBS, which will react much faster. A specimen which I have in my laboratory weighs 0.75 mg/cm^2 . The lag conditions for this material are calculated in table 5.

Table 5

h	T	P	lag at 50% RH			lag at 20% RH			lag at 10% RH		
			int.	boun.	total	int.	boun.	total	int.	boun.	total
0	276	1	0.01	0.02	0.03	0.07	0.03	0.1	0.16	0.05	0.2
3	256	0.7	0.03	0.05	0.08	0.13	0.07	0.2	0.3	0.13	0.4
6	240	0.5	0.06	0.11	0.2	0.28	0.17	0.45	0.7	0.3	1.0
9	222	0.3	0.13	0.35	0.5	0.7	0.55	1.25	1.6	1.0	2.6
12	213	0.2	0.27	0.6	0.9	1.3	0.9	2.2	3.2	1.6	4.8

Table 5 shows that GBS of single thickness ought to give quite good results at temperatures as low as -50°C, with a certain reservation at the lowest relative humidity. It should be pointed out, however, that the diffusion process is only slowed down when the indication of the GBS instrument has already reached the neighbourhood of this relative humidity.

To give an example. A humidity change at 222° from 15% RH to 5% RH will result in a change of indication to about 10% RH only within two and a half minutes. But a sudden change of humidity from 55% RH to 5% RH will reach an indication of 15% within the same time. This shows that the limitations, while affecting the accuracy, are not entirely prohibitive.

The reduced strength of this thinner material will affect the instrumental design in so far, as it may become necessary to use several strips of GBS in parallel to operate the indicating mechanism. (Provision for this has already been made in the design which I sent to Boscombe Down where room has been left to fit three GBS strips in parallel). It may also be possible that the thinner material may not stand up to an air speed of 100 miles/hour, but a reduction of the air flow to 75 or even 50 miles/hour will not seriously affect the values in table 5, as only the boundary component of the lag will be altered and that with $\sqrt{u}$.

September 28, 1943.

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

M.R.P. 143
7th October 1943

Extract from Quarterly Report on work done for the Meteological Office during July to September 1941 at the Science Laboratories of the Durham Colleges, Durham.

by E. Gluckauf.

A. Goldbeater Skin Hygrometer.

Calibration curves have been taken for relative humidities from 0 to 100% at temperatures ranging from +18°C to -41°C. These show that, at temperatures below zero, the GBSH indicates relative humidity with respect to supercooled water, similar to the hair hygrometer. Calibrations at room temperature can, therefore, be applied without correction to all temperatures. (See Appendix, tables 1 and 2.)

Like the hair hygrometer, the GBSH also suffers from a hysteresis effect after exposure to relative humidities lower than 30%. (See appendix, Figure 1.) This hysteresis effect takes place also at low temperatures; curves, similar to Fig.1 have been observed at -21°C and -28°C. The effect of exposure to low humidities disappears again at humidities higher that 60%RH. with respect to water, and does this also at very low temperatures.

The lag of the GBSH has been tested at temperatures between +18°C and -68°C, using air velocities from 1.2 to 6 metres/second. In figure 2 the logarithms of the half-way times (τ) are plotted against the reciprocal absolute temperatures (T), resulting in straight lines for the various air velocities. It is to be noted that the temperature coefficient of the response constant ($k = \frac{\log 2}{\tau}$) is somewhat smaller than that of the saturation pressure of liquid water.

Figure 3 shows the esponse constant k for small changes at various humidities; at 18°C both at low and high humidities the half-way times are longer than at medium humidities. This is quite different from the behaviour of the hair hygrometer, for which, as shown in previous work, the half-way times decrease throughout towards higher humidities. Figure 4 shows that this phenomenon is substantially the same at lower temperatures (-21°C). The maximum of the response constant k (minimum of τ.) again lies at about 55% RH with respect to water, while the highest rel. humidities, at which k would be very small, are unobtainable on account of ice saturation. It also appears that movements towards increased humidities are somewhat faster than those in the opposite direction, for which an explanation is so far missing.

The observed data make it clear that, at air velocities up to 6 m/s, internal diffusion in the gold beater skin is not the lag-determining factor. The question whether the actual arrival of water vapour at the boundary layer, or the transfer through the boundary layer, or both, are dominant factors, cannot be safely decided at present. However, it would appear, - as explained in detail in the appendix - that at the lowest temperatures the former, and at the highest temperatures the latter cause, are the decisive factors. It is intended to attempt the solution of this question, which is of interest on account of the influence of lower pressures, by direct experiment under reduced pressure. Another practical point which also requires attention is, that under experimental conditions, - flow of air through a narrow channel along the length of the GBskin (95 mm long), - the length of the skin may have an influence on the lag, especially at very low temperatures where the arrival of water vapour seems to be the dominant factor.

Appendix.

1. Calibration of GOLDBEATER SKIN HYGROMETER at -21°C.

A. %RH of air used, rel. to ice	B. %RH indicated acc. to room temperature calibration.	C. %RH of air used, relative to supercooled water.
100	81, 78.5, 79	81.0
85	70	69.8
67.5	54, 55	54.6
57.5	46	46.6
52.5	41.5	42.5
42.5	34, 33	34.4
32.5	26, 27.5, 27	26.3
15	11.5, 12.5	12.1

II. Calibration at temperatures below 0°C with air saturated over ice.

Temperature °C	%RH indicated acc. to room temperature calibration.	%RH of air used, relative to supercooled water.
+ 0.8	100	100
-2.3	97.5	97.9
-6.5	93.5	94.0
-10.5	90.5	90.5
-15.5	85.1	86.1
-19.2	82.6	82.9
-22.0	80.4	80.7
-25.6	77.5	77.7
-30.0	73.7	74.2
-33.5	72.5	71.6
-36.5	70.0	69.6
-39.5	67.1	67.5
-42.5	65.7	65.6
-65	53	52.6

Remarks on Figure 1. The hysteresis effects are probably produced by a change of modification of the keratine. There seem to be at least two of these, stable at high and low humidities respectively, one of which is represented by the linear part of the calibration curve between 100 and 50%RH, and which becomes unstable below 50%RH. That there is a change of melification is clearly shown by a number of lag curves obtained for changes from 52% to lower RH. In one case, the GBSH, after a quick change to 15% RH had come practically to a standstill at a point of 4.40% contraction (in brackets on the figure); then, suddenly it began to contract further until it reached a point of 5.65% contr., lying on the main curve. This observation may be of practical importance for the extrapolation of actual humidities from flight curves, as this would lead to a point on the dotted line, and not to the normal calibration value.

Remarks on figure 2.

Of the air velocity ranges (u) only the 1.2 and 1.7 m/sec lines are well established, while the 3.3 and 5-6 m/s lines suffer from considerable fluctuations of their points. The increase in response between 1.2 and 1.7 m/s is with $u^{1.1}$ or approximately proportional to u, while between 1.7 and 6 m/s the increase goes approximately with $u^{0.6}$. This behaviour can be easily explained, as, at the very lowest velocities the arrival of moisture must be the dominant factor which would make the response proportional to u. At higher air velocities, however, when sufficient water arrives at the boundary layer, the response will be governed by diffusion through the latter. The thickness of the boundary layer and, consequently, the half-way time may be expected to vary inversely to $u^{0.5}$. Only at those air velocities where the boundary layer is the lag-determining factor, can reduction of barometric pressure be expected to lead to an increase in response.

However, it is very doubtful, whether the boundary layer comes into play at the lowest temperatures at the air velocities used. The amount of water required for GBS for a change of 10%RH (water) at medium humidities can be expected to be about 2.0% of the weight of the skin. Those used weighed 28 mgr., requiring 0.56 mgr. of water, or 0.28 mgr. up to the half-way time. At -55°C one litre of air contains about 0.37 mgr water at water saturation. If the air could give up all its surplus, i.e.:10% of 0.037 mgr, 76 litres of air would be required. Actually the amount it can supply up to the half-way time is only 3/4 of this amount, which brings the total of air required to 101 litres of air. The half-way time found was 14 minutes, using 11.3 litres of air per minute, or a total of 158 litres up to half-way time. This is only 57% more than the theoretically required amount, and it would be surprising, if under these conditions the response were not governed by the actual supply of water.

Correction: These air velocities have been measured at room temperature. At lower temperatures they would be correspondingly lower.

Remarks on figures 3, 4 and 5.

The difference of these curves as compared with those of the hair hygrometer, is probably due to the different mechanisms dominating the speed of response. In the hair it is internal diffusion, for which the speed increases with high humidities. InGBS, assuming the same diffusion coefficient for water in the keratine, internal diffusion should be 20 to 30 times faster than in hair. Under these conditions the lag is dominated by either the supply of water vapour or by diffusion through the boundary layer. Either of these would make the response proportional to d(RH)/dW, where W is the amount of water taken up by GBS for a given RH with resp. to water. Figure 5 shows the water content of wool (King and Cassie, Trans. Farad. Soc. 36, 449, 1940), a chemically very similar substance. While no quantitative agreement can be expected, it is clear from Fig. 5 that d(RH)/dW has, like the response constant, a maximum at medium humidities (with respect to water).

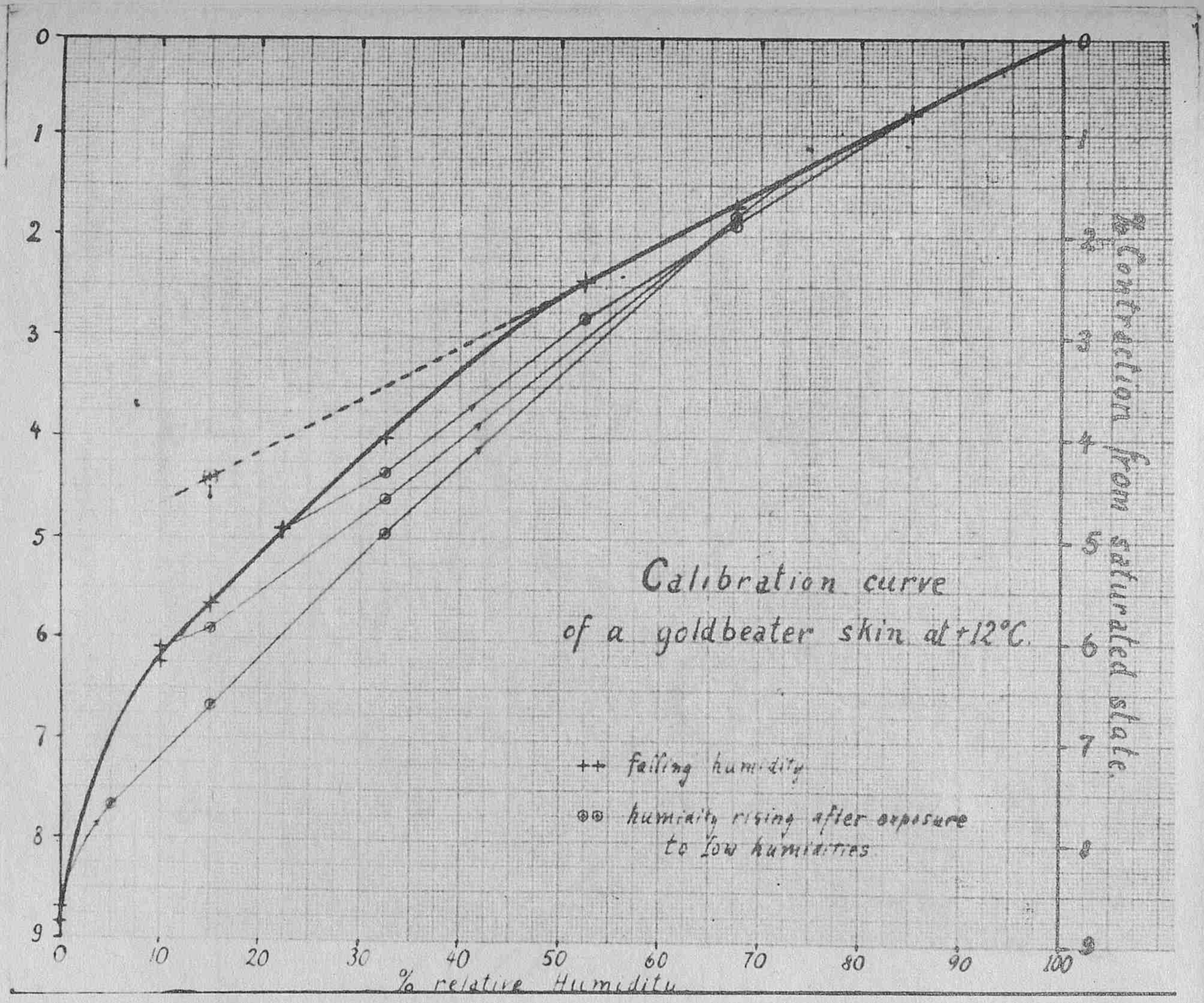
Calibration curve
of a goldbeater skin at +12°C.
++ falling humidity
⊕⊕ humidity rising after exposure to low humidities
% Contraction from saturated state.
% relative Humidity
0
1
2
3
4
5
6
7
8
9
0
10
20
30
40
50
60
70
80
90
100

FIGURE I

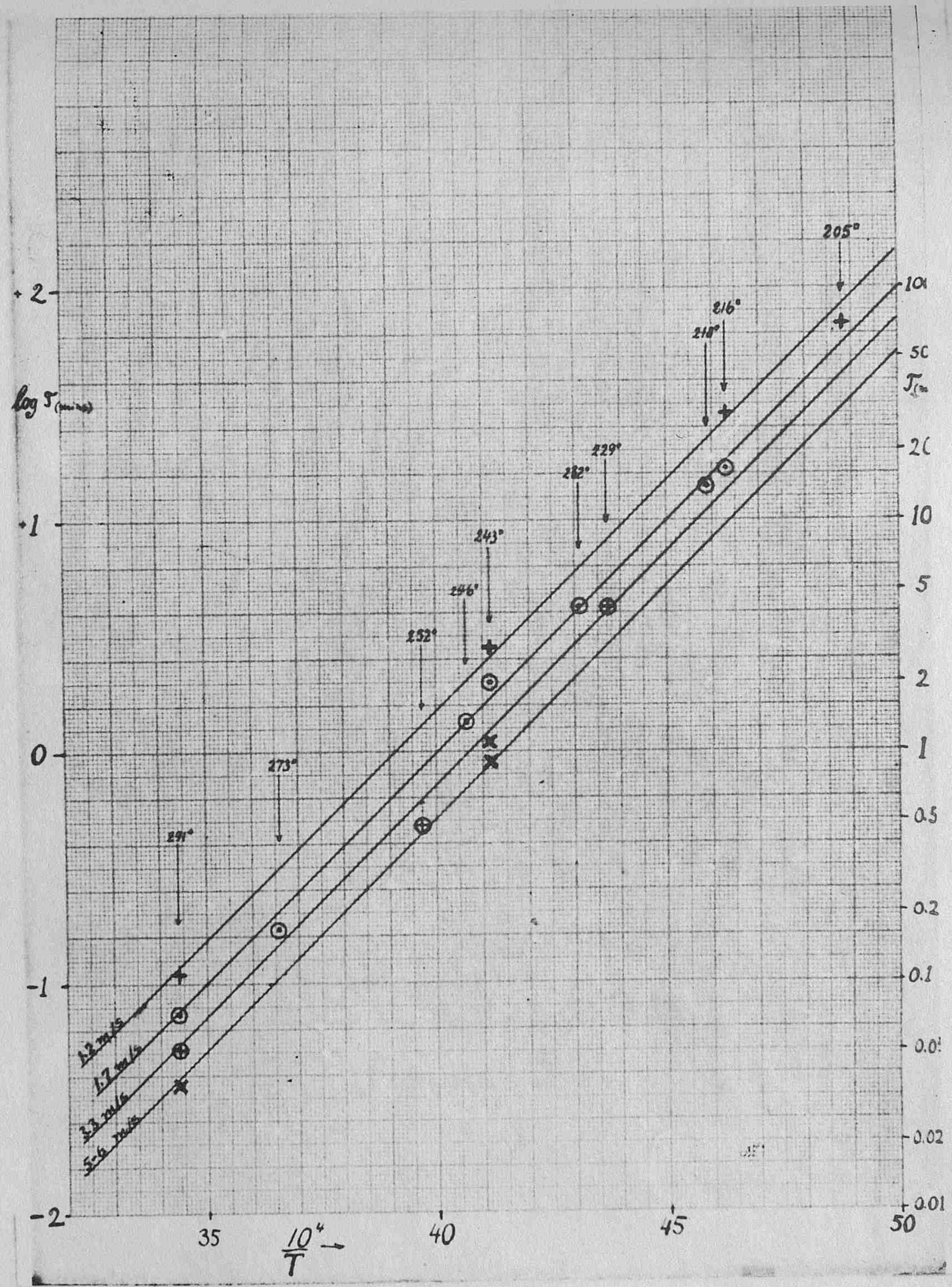205°
216°
218°
229°
232°
243°
246°
252°
273°
291°
+2
+1
0
-1
-2
log τ (mins)
1.2 m/s
1.7 m/s
3.3 m/s
5-6 m/s
35
40
45
50
10⁴/T →
100
50
τ (m
20
10
5
2
1
0.5
0.2
0.1
0.05
0.02
0.01

FIGURE 2

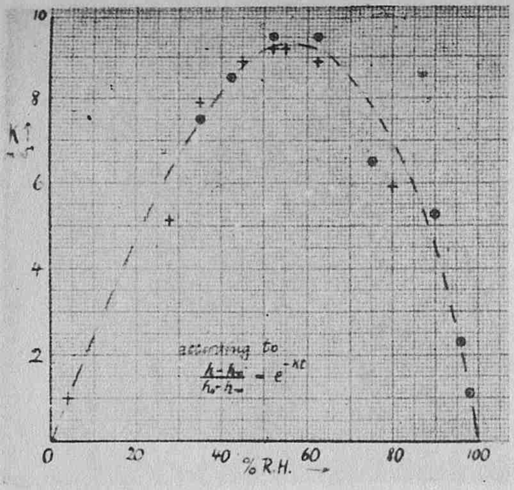

FIGURE 3

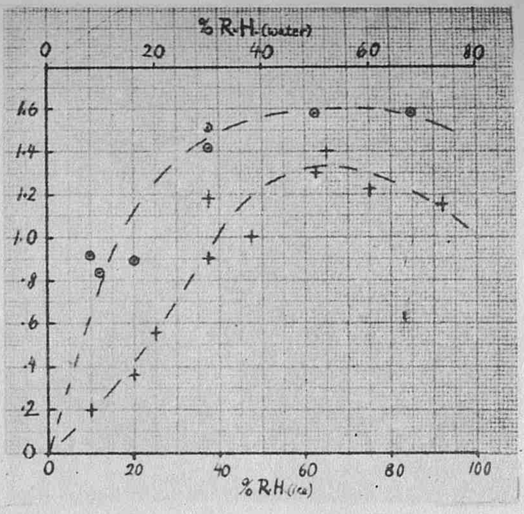

FIGURE 4

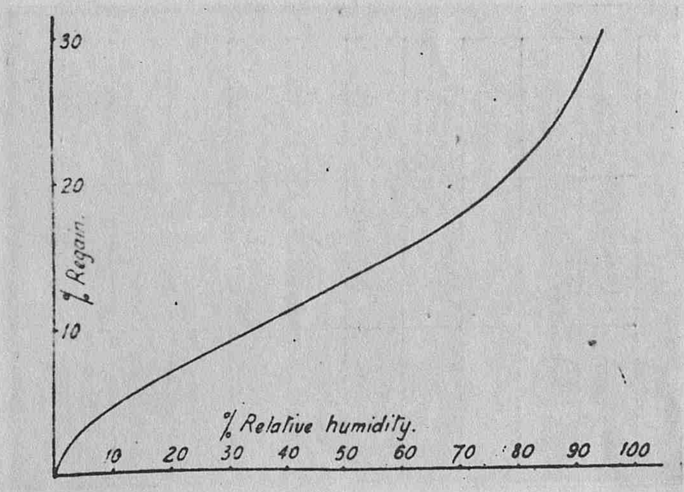

FIGURE 5

Navaer 50-1T-28

M.R.P.147
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METEOROLOGICAL RESEARCH COMMITTEE

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GEOSTROPHIC AND NON-GEOSTROPHIC MOTION

by
Sverre Petterssen

(June 1943)

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INTRODUCTION.

The aim of this investigation is to contribute to the development of methods by the aid of which non-geostrophic motion can be analysed and the accelerations assessed.

A direct evaluation of the geostrophic departures through comparing the actual wind with the geostrophic wind is not practicable, for the geostrophic departures are frequently smaller than the errors in the observations. On the other hand, evidence is gradually accumulating in support of the view that the degree of geostrophic balance, or the lack of such balance, is revealed to the analyst through the patterns of pressure and density that the upper air charts present. It is, therefore of importance to investigate the dynamical significance of these patterns.

As a logical approach to the solution of this problem we shall first explore the exact dynamical properties of, and criteria for, geostrophic balance, so as to enable the synoptic meteorologist to recognise a geostrophic case, when one occurs. The criteria thus obtained are such that their non-fulfilment can be ascertained by the aid of synoptic upper air charts and a measure of the unbalanced part of the motion obtained.

The contents of this report are limited to the derivation of the various criteria for geostrophic balance together with an exposition of some of the most important synoptic symptons of unbalanced motion.

Specific types of unbalanced motion as well as possible practical applications will be discussed in later reports.

I. THE EQUATIONS OF MOTION.

1. The System of Co-ordinates.

The system of co-ordinates most suited for a study of atmospheric motion on a large scale is a spherical system (r, φ, λ) in which distance (r) from the centre of the earth, latitude (φ) and longitude (λ) are the co-ordinates. In this system the gravity force is everywhere perpendicular to the $\varphi\lambda$- surface, and in horizontal motion

$$\frac{dr}{dt} = 0.$$

In many problems it is convenient to use a rectangular system of co-ordinates (x, y, z) with the xy-plane coinciding with the horizon plane, and the z-axis positive upward. In this system, gravity is perpendicular to the xy-plane only along the z-axis (i.e. x = y = 0).

If $\frac{dz}{dt} = 0$ is taken to define horizontal motion, it is evident that

$$\frac{dz}{dt} = \frac{dr}{dt}$$

only along the z-axis (i.e. x = y = 0).

Furthermore, if ω is the angular velocity of the earth's rotation, the horizon plane in latitude φ has a rotation equal to

$$\omega \sin\varphi$$

which is a parameter that must be regarded as a constant in x and y once the system of co-ordinates has been chosen. The equations of motion as normally written down for rectangular systems of co-ordinates, are equivalent to those that hold in a sperical system within an area around the origin that is sufficiently small for the components of gravity in the xy-plane to be disregarded. Within this area, φ is to be regarded as a constant parameter.

In this investigation, the rectangular system of co-ordinates will be used, and this system is adequate since the discussion is limited to the differential mechanics of atmospheric motion.

2. The Equations of Motion.

In a right-handed rectangular system of co-ordinates with the x-axis toward the east, the y-axis toward the north, and the z-axis along the vertical, the equations of frictionless motion may be written

$$\left.\begin{aligned} \rho_s \dot{u} &= -\frac{\partial p}{\partial x} + \rho_s v\, 2\omega \sin\varphi - \rho_s w\, 2\omega \cos\varphi \\ \rho_s \dot{v} &= -\frac{\partial p}{\partial y} - \rho_s u\, 2\omega \sin\varphi \\ \rho_s \dot{w} &= -\frac{\partial p}{\partial z} - \rho_s g + \rho_s u\, 2\omega \cos\varphi \end{aligned}\right\} \qquad (1)$$

where p = atmospheric pressure

ρ_s = density ~~(all ρ_s are marked in red)~~

u, v, w = the components of the velocity along the co-ordinate axes

$\dot{u}, \dot{v}, \dot{w}$ = the components of the acceleration along the co-ordinate axes (the dot symbol being used to denote differentiation with regard to time following an individual unit of air).

g = the acceleration of gravity

ω = the angular velocity of the earth's rotation

φ = latitude.

It is sometimes convenient to introduce specific volume (α): instead of density. The equation may then be written

$$\left.\begin{aligned} \dot{u} &= -\alpha\frac{\partial p}{\partial x} + v2\omega\sin\varphi - W2\omega\cos\varphi \\ \dot{v} &= -\alpha\frac{\partial p}{\partial y} - u2\omega\sin\varphi \\ \dot{w} &= -\alpha\frac{\partial p}{\partial z} - g + u2\omega\cos\varphi \end{aligned}\right\} \quad (2)$$

Either of the above sets of equations are convenient to use in connection with charts showing the distribution of the variables in horizontal planes.

In upper air analyses it is found preferable to construct maps showing the distribution of the variables in standard surfaces of constant pressure. The most convenient form for the equations of motion is then obtained by eliminating α between the equations (2). This gives:

$$\left.\begin{aligned} \dot{u} &= -g\tan A\left(1+\frac{\dot{w}-2\omega\cos\varphi u}{g}\right) + 2\omega(v\sin\varphi - w\cos\varphi) \\ \dot{v} &= -g\tan B\left(1+\frac{\dot{w}-2\omega\cos\varphi u}{g}\right) - 2\omega u\sin\varphi \end{aligned}\right\} \quad (3)$$

Here

$$\tan A = \frac{dz}{dx} = -\frac{\partial p}{\partial x}\Big/\frac{\partial p}{\partial z}, \qquad \tan B = \frac{dz}{dy} = -\frac{\partial p}{\partial y}\Big/\frac{\partial p}{\partial z}$$

Thus, tan A and tan B denote the slope of the pressure surface in the direction of the horizontal co-ordinate axes.

The three sets of equations (1), (2) and (3) are equivalent and the choice between them is entirely a matter of convenience. In the following sections, the restrictions imposed by such assumptions as it has ~~have~~ become customary to make in meteorology will be discussed in considerable detail, particularly with a view to deriving precise criteria for geostrophic motion and such indications of non-geostrophic motion as can be ascertained by the aid of synoptic charts.

II. RESTRICTIONS IMPOSED BY ASSUMING HYDROSTATIC EQUILIBRIUM.

3. Hydrostatic Equilibrium.

If hydrostatic equilibrium is assumed

$$-\frac{\partial p}{\partial z} = \rho g$$

and the third of eqs.(1) reduces to

$$\dot{w} = 2\omega\cos\varphi \cdot u$$

In the middle latitudes

$$2\omega\cos\varphi \approx 10^{-4}$$

In the free atmosphere, u is usually in excess of 10m/sec. and sometimes surpasses 50 m/sec. Even with a low value for u

$$2\omega\cos\varphi \cdot u > 10^{-3}$$

If the vertical acceleration were to balance the rotational term a vertical velocity of at least 1 m/sec. would result in less than 15 minutes. In the large scale currents of the atmosphere vertical velocities in excess of 0.05 m/sec. are rare, and the vertical accelerations such that it takes hours to produce these velocities. A high estimate of the magnitude of the large scale vertical acceleration is therefore

$$\dot{w} \approx 10^{-5}$$

whereas a low estimate of the rotational term is

$$2\omega\cos\varphi\, u \approx 10^{-3}$$

We may, therefore, neglect the vertical acceleration against the term containing u. The third of eqs.(1) then reduces to:

$$0 = \frac{\partial p}{\partial z} - \rho_s(g - 2\omega \cos\varphi\, u) \qquad (4)$$

and the corresponding eqs.(2) and (3) are similarly simplified.

4. Dynamic Pressure.

If P denotes the static pressure

$$\frac{\partial p}{\partial z} = -\rho_s g$$

and

$$\Delta p = P - p$$

expresses the dynamic part of the pressure. Eq.(4) may then be written

$$0 = \frac{\partial \Delta p}{\partial z} + \rho_s u\, 2\omega \cos\varphi \qquad (5)$$

An estimate of the magnitude of the vertical gradient of the dynamic pressure can be obtained by assuming that the motion along the parallel of latitude (i.e.u) is geostrophic. Then

$$2\omega \sin\varphi \cdot \rho_s u = -\frac{\partial p}{\partial y}$$

and eq.(5) becomes

$$\frac{\partial \Delta p}{\partial z} = \frac{\partial p}{\partial y} \operatorname{Cotan}\varphi$$

Thus, for example, in latitude 45° the vertical gradient of the dynamic pressure is exactly equal to the horizontal pressure gradient along the meridian.

Integrating eq. (5) from $z = z_0$ to $z = \infty$, we obtain

$$\Delta p = 2\omega \cos\varphi \int_{z_0}^{\infty} \rho_s u\, dz$$

and the dynamic portion of the pressure in a point is proportional to the specific momentum due to the zonal circulation of the atmosphere above that point.

To obtain an estimate of the dynamic pressure we may assume u to be constant along the vertical up to such heights where the density is negligible. Then

$$\Delta p = \frac{2\omega\cos\varphi}{g} \cdot u \cdot \int_{z_0}^{\infty} \rho_s g\, dz = \frac{2\omega\cos\varphi\, u}{g} \cdot P$$

If $P = 1000$ mb. and $\varphi = 45°$

$$\Delta p = 0{\cdot}01\, u \quad (\text{mb.})$$

and even if the zonal component of the mean wind were as large as 50 m/sec., the dynamic portion of the pressure would only amount to 0.5 mb. which is well below the errors due to the imperfection of the instruments. Therefore, as far as height computations are concerned it is entirely satisfactory to assume hydrostatic equilibrium.

5. Circulation.

Although it is satisfactory to compute the heights of pressure surfaces on the assumption of hydrostatic equilibrium, the deviation from the hydrostatic conditions may exert an influence on the creation and destruction of circulation. From the complete equations (2) we obtain for the components of the curl of the acceleration:

$$\operatorname{curl}_x \dot{V} = -\left(\frac{\partial \alpha}{\partial y} \cdot \frac{\partial p}{\partial z} - \frac{\partial \alpha}{\partial z} \cdot \frac{\partial p}{\partial y}\right) + 2\omega\left(\frac{\partial u}{\partial y} \cos\varphi + \frac{\partial u}{\partial z} \sin\varphi\right)$$

$$\operatorname{curl}_y \dot{V} = -\left(\frac{\partial \alpha}{\partial z} \cdot \frac{\partial p}{\partial x} - \frac{\partial \alpha}{\partial x} \cdot \frac{\partial p}{\partial z}\right) + 2\omega\left(\frac{\partial v}{\partial z} \sin\varphi - \frac{\partial u}{\partial x} \cos\varphi\right) - \frac{\partial w}{\partial z} \cdot 2\omega \cos\varphi$$

$$\operatorname{curl}_z \dot{V} = -\left(\frac{\partial \alpha}{\partial x} \frac{\partial p}{\partial y} - \frac{\partial \alpha}{\partial y} \cdot \frac{\partial p}{\partial x}\right) - 2\omega \sin\varphi \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) + 2\omega \cos\varphi \frac{\partial w}{\partial y}$$

If we assume hydrostatic equilibrium, the terms

$$2\omega\cos\varphi\frac{\partial u}{\partial y} \quad \text{and} \quad 2\omega\cos\varphi\frac{\partial u}{\partial x}$$

would disappear from the first and second of the above equations, whereas, the third would be unaffected. The assumption of hydrostatic equilibrium does, therefore, not directly affect the formation of vorticity in the horizontal plane. The formation of vorticity in either of the vertical planes is affected only to the extent that the horizontal shear terms (i.e. $\frac{\partial u}{\partial x}$ and $\frac{\partial u}{\partial y}$) are neglected in comparison with the vertical shear terms (i.e. $\partial u/\partial z$ and $\partial v/\partial z$). Normally, the vertical shear terms are much larger than the corresponding horizontal shear terms. Except in the frontal zones, where the horizontal shear is important, it appears that no essential feature of the motion is omitted through the assumption of hydrostatic equilibrium.

III. GEOSTROPHIC BALANCE.

6. Geostrophic Motion.

It is customary to define geostrophic motion as non-accelerated motion in which the horizontal pressure force is exactly balanced by the Coriolis force resulting from the horizontal motion. Thus, one of the conditions that must be satisfied is

$$\dot{u} = \dot{v} = \dot{w} = 0$$

and eqs.(1) reduce to

$$0 = -\frac{\partial p}{\partial x} + 2\omega\sin\varphi\cdot\rho_s v - 2\omega\cos\varphi\cdot\rho_s w \qquad (6)$$

$$0 = -\frac{\partial p}{\partial y} - 2\omega\sin\varphi\cdot\rho_s u \qquad (7)$$

$$0 = -\frac{\partial p}{\partial z} - \rho_s g + 2\omega\cos\varphi\cdot\rho_s u \qquad (8)$$

It should now be ascertained that the above equations are mutually consistent, consistent with the definition and in agreement with the equation of continuity. The latter question will be descussed in para.12.

Multiplying the above equations by u, v and w, and adding, we obtain

$$\frac{\partial p}{\partial x}u + \frac{\partial p}{\partial y}v = -\left(\frac{\partial p}{\partial z} + \rho_s g\right)w$$

and the work done by the horizontal pressure force must be balanced by the work done by the vertical gradient of the dynamic pressure.

The definition of geostrophic balance requires that the horizontal motion is parallel to the isobars, or

$$\frac{\partial p}{\partial x}u + \frac{\partial p}{\partial y}v = 0$$

This is possible either when

$$w = 0$$

or when

$$\frac{\partial p}{\partial z} + \rho_s g = 0$$

This latter condition is true only when there is no motion along the parallels of latitude (i.e. $u = 0$). This special case could not be geostrophic since the isobars would then be along, and converge with, the meridians, thus giving rise to accelerations. Therefore, the motion can be non-accelerated and along the isobars only when the motion is strictly horizontal.

The fundamental properties of geostrophic motion is, therefore, expressed by

$$\boxed{\dot{u} = \dot{v} = \dot{w} = w = 0} \qquad (9)$$

7. Synoptic Formulation of the Problem.

It is important to note that all that is implied in eq.(9) is that the accelerations and the vertical velocity vanish at a point (x,y,z) at the instant considered; nothing is stated as regards the extension of the geostrophic balance in space and time. Eq.(9), therefore, specifies the conditions that characterize singular or accidental geostrophic balance.

From the point of view of synoptic meteorology, we are not particularly interested in singular, or accidental, geostrophic balance, such as may occur when a parcel of air passes from a positive to a negative acceleration, or vice versa for the meshes in our network of stations are such that the geostrophic fulfilment in a singular point cannot be ascertained.

The type of geostrophic wind of interest in synoptic meteorology must be such that if the motion is geostrophic at, say, Victory House, Kingsway, it must be geostrophic also in Bayswater, and, furthermore, it must be geostrophic not only in a level z = constant, but also within a layer of finite thickness around that level. This is the same as saying that the geostrophic balance must obtain not only at a singular point (x,y,z) but also at any neighbouring point (x + dx, y + dy, z + dz). This requirement may be expressed by:

$$\boxed{\frac{\partial}{\partial x} = \frac{\partial}{\partial y} = \frac{\partial}{\partial z} = 0} \qquad (10)$$

when applied to $\dot{u}$, $\dot{v}$, $\dot{w}$ and w.

In the following, this type of geostrophic motion will be discussed in detail, particularly with a view to discovering its exact dynamical properties and the synoptic criteria for its existence. This type of motion, will, when necessary, be referred to as general geostrophic motion to distinguish it from the singular type.

8. Equations of Geostrophic Motion.

Introducing eqs.(9) into eqs.(1), we obtain:

$$0 = -\frac{\partial p}{\partial x} + 2\omega \sin\varphi \cdot \rho_s v$$

$$0 = -\frac{\partial p}{\partial y} - 2\omega \sin\varphi \, \rho_s u$$

$$0 = -\frac{\partial p}{\partial z} - \rho_s g + 2\omega \cos\varphi \, \rho_s u$$

These equations represent an inertia motion since no work is done by any of the forces involved.

It is convenient (although not necessary) and sufficiently accurate to neglect the rotational term in the third of the above equations. With this simplification we obtain from eqs.(1), (2) and (3) the following equivalent sets of equations:

$$0 = -\frac{\partial p}{\partial x} + 2\omega \sin\varphi \, \rho_s v \qquad (11)$$

$$0 = -\frac{\partial p}{\partial y} - 2\omega \sin\varphi \, \rho_s u \qquad (12)$$

$$0 = -\frac{\partial p}{\partial z} - \rho_s g \qquad (13)$$

$$0 = -\alpha \frac{\partial p}{\partial x} + 2\omega \sin\varphi\, v \qquad (14)$$

[redacted]

$$0 = -\alpha \frac{\partial p}{\partial y} - 2\omega \sin\varphi\, u \qquad (15)$$

$$0 = -\alpha \frac{\partial p}{\partial z} - g \qquad (16)$$

$$0 = -g \tan A + 2\omega \sin\varphi\, v \qquad (17)$$

$$0 = -g \tan B - 2\omega \sin\varphi\, u \qquad (18)$$

It should be noted that the complete equations of motion (1 - 3) apply when the x-axis is chosen along a parallel of latitude. The simplified equations (11 - 18) hold in any rectangular system with the z-axis along the vertical. We can therefore, when using the latter equations, choose the direction of the horizontal axes according to convenience. In the following discussion it is often convenient to choose the x-axis along the isobar, such that $\frac{\partial p}{\partial y} < 0$.

9. Vertical Variation of the Horizontal Pressure Gradient.

Differentiating eq.(13) with regard to x and y, we obtain

$$-\frac{\partial}{\partial x}\left(\frac{\partial p}{\partial z}\right) = -\frac{\partial}{\partial z}\left(\frac{\partial p}{\partial x}\right) = g\,\frac{\partial \rho_s}{\partial x}$$
$$-\frac{\partial}{\partial y}\left(\frac{\partial p}{\partial z}\right) = -\frac{\partial}{\partial z}\left(\frac{\partial p}{\partial y}\right) = g\,\frac{\partial \rho_s}{\partial y} \qquad (19)$$

Thus, if the density varies along the isobars at the level z_0, the isobars will rotate with increasing, or decreasing, elevation.

It is well to bear in mind that this well-known result, which is derived from the hydrostatic equation, holds in general motion. Therefore, it does not necessarily follow that geostrophic motion is possible when the density varies along the isobars. We shall return to this question later.

10. Curved Isobars.

Differentiating eq.(11) partially with respect to t and x, we obtain:

$$\frac{\partial^2 p}{\partial x \partial t} = 2\omega \sin\varphi \frac{\partial \rho_s}{\partial t} v + 2\omega \sin\varphi \frac{\partial v}{\partial t}$$
$$\frac{\partial^2 p}{\partial x^2} = 2\omega \sin\varphi \frac{\partial \rho_s}{\partial x} v + 2\omega \sin\varphi \frac{\partial v}{\partial x}$$

Choosing the x-axis along the isobar, v vanishes, and when the isobar is curved

$$\frac{\partial^2 p}{\partial x^2} \gtrless 0$$

Multiplying the last equation by u, and adding:

$$\frac{\partial^2 p}{\partial x \partial t} + u \frac{\partial^2 p}{\partial x^2} = 0 \qquad (20)$$

since

$$\frac{\partial v}{\partial t} + \frac{\partial v}{\partial x} u = \dot{v} = 0$$

Now, the velocity of an element of a curved isobar in the direction tangent to itself is expressed by*;

$$c_x = -\frac{\partial^2 p}{\partial x \partial t} \Big/ \frac{\partial^2 p}{\partial x^2} \qquad (21)$$

*For Footnote see Page 8.

This substituted into eq.(20) gives :

$$\frac{\partial^2 p}{\partial x^2}\left(u - C_x\right) = 0.$$

and since, by hypothesis $\frac{\partial^2 p}{\partial x^2} \gtrless 0,$ $\quad u = C_x$

Geostrophic motion in curved isobars is therefore possible only at points where the velocity is equal to the speed of the pressure system. As an example, we consider Fig.1 in which the air motion is assumed to be geostrophic while the pressure trough moves with the speed C_x. It is evident that the wind velocity cannot be along the isobars and also equal to C_x, except on the trough line, and even there it can at most be so at a singular point. To obtain general geostrophic motion, therefore, it is necessary that

$$\boxed{\frac{\partial^2 p}{\partial x^2} = 0} \tag{22}$$

or, the isobar considered must be a straight line.

For this condition to hold also along the line element dy, it is necessary that

$$\frac{\partial}{\partial y}\left(\frac{\partial^2 p}{\partial x^2}\right) = \frac{\partial}{\partial x}\left(\frac{\partial^2 x}{\partial x \partial y}\right) = 0$$

There are here two possibilities, viz,

$$\frac{\partial^2 p}{\partial x \partial y} = a \text{ constant} \gtrless 0$$

or

$$\boxed{\frac{\partial^2 p}{\partial x \partial y} = 0} \tag{23}$$

The first alternative is compatible with eq.(22) only in a col, that is $\partial p/\partial y = 0$, which means that the geostrophic motion vanishes. It follows then that eq.(23) must hold, and general geostrophic motion is possible only when the isobars are straight and parallel.

11. Straight Isobars.

Choosing the x-axis along an isobar,

$$\frac{\partial p}{\partial x} = 0 \quad \text{and} \quad v = 0 \tag{24}$$

Furthermore,

$$\left.\begin{aligned} \dot{u} &= \frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} \\ \dot{v} &= \frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} \end{aligned}\right\} \tag{25}$$

Differentiating each of eqs.(11), (12) and (13) with regard to x, y, z, and t, and taking account of eqs.(20), (22), (23), (24) and (25), we obtain the following relations that must hold in general geostrophic motion:

$$\frac{\partial v}{\partial t} = \frac{\partial v}{\partial x} = \frac{\partial v}{\partial y} = 0 \tag{26}$$

$$\frac{\partial^2 p}{\partial x \partial t} = 0 \tag{27}$$

$$\frac{\partial^2 p}{\partial x \partial z} = -g\frac{\partial \rho_s}{\partial x} = 2\omega \sin\varphi\, \rho_s \frac{\partial v}{\partial z} \tag{28}$$

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*Sverre Petterssen: "Kinematical and Dynamical Properties of the Field of Pressure with Application to Weather Forecasting." Geof.. Pub. Vol. X, No.2, 1933.

$$\rho_s \frac{\partial u}{\partial x} = -u \frac{\partial \rho_s}{\partial x} \qquad (29)$$

$$\frac{\partial^2 p}{\partial y^2} = -2\omega \sin\varphi \left(\rho_s \frac{\partial u}{\partial y} + u \frac{\partial \rho_s}{\partial y} \right) \qquad (30)$$

$$\frac{\partial^2 p}{\partial y \partial z} = -g \frac{\partial \rho_s}{\partial y} = -2\omega \sin\varphi \left(\rho \frac{\partial u}{\partial z} + u \frac{\partial \rho_s}{\partial z} \right) \qquad (31)$$

$$\frac{\partial^2 p}{\partial y \partial t} = -2\omega \sin\varphi \left(\rho_s \frac{\partial u}{\partial t} + u \frac{\partial \rho_s}{\partial t} \right) \qquad (32)$$

$$\frac{\partial}{\partial t}\left(\frac{\partial p}{\partial y} \right) = -2\omega \sin\varphi \, u \frac{\partial \rho_s}{\partial t} \qquad (33)$$

$$\frac{\partial^2 p}{\partial z \partial t} = -g \frac{\partial \rho_s}{\partial t} \qquad (34)$$

These relations, which were derived from the equation of hydrostatic equilibrium and the equations of geostrophic motion, must satisfy the equation of continuity.

12. The Equation of Continuity.

In accelerated motion the equation of continuity may be written:

$$\frac{\partial \rho_s}{\partial t} = \frac{\partial(\rho_s u)}{\partial x} + \frac{\partial(\rho_s v)}{\partial y} + \frac{\partial(\rho_s w)}{\partial z}$$

Suppose now that

$$\dot{u} = \dot{v} = \dot{w} = \frac{\partial w}{\partial z} = 0$$

at a point, and that the accelerations in the vicinity of that point are controlled by a potential. Since

$$\text{curl}\,(\text{grad}) = 0$$

we obtain from first two eqs.(1)

$$\frac{\partial \rho_s u}{\partial x} + \frac{\partial \rho_s v}{\partial y} = 0$$

and the equation of continuity reduces to

$$\frac{\partial \rho_s}{\partial t} = 0 \qquad (35)$$

This combined with the hydrostatic equation gives

$$\frac{\partial^2 p}{\partial z \partial t} = 0 \qquad (36)$$

Now, if the acceleration potential reduces to a constant in x and y, the accelerations vanish not only at a singular point, but also in any neighbouring point, and we have general geostrophic motion. The condition of continuity then requires that not only eqs.(35) and (36) are satisfied but also that

$$\frac{\partial^2 \rho_s}{\partial x \partial t} = \frac{\partial^2 \rho_s}{\partial y \partial t} = 0$$

or that

$$\frac{\partial}{\partial z}\left(\frac{\partial^2 p}{\partial x \partial t} \right) = \frac{\partial}{\partial z}\left(\frac{\partial^2 p}{\partial y \partial t} \right) = 0 \qquad (37)$$

Eqs.(27) and (36) state that the components of the isallobaric gradient along the x-axis (i.e. along the isobar) and along the z-axis vanish. Therefore if an isallobaric gradient is compatible with the geostrophic motion, it must be horizontal and perpendicular to the isobars (i.e. along the y-axis).

Now, eq.(37) states that the isallobaric gradient along the y-axis must not vary with elevation (z). On the other hand eqs.(19) (which hold in general motion) state that if the density varies along the isobars, the isobars will rotate with increasing or decreasing elevation.

Obviously, the horizontal isallobaric gradient cannot be constant along the vertical and also perpendicular to the isobars if the latter rotate with elevation. From this it follows that either

$$\frac{\partial \rho_s}{\partial x}=0 \qquad \text{(alternative 1)}$$

or

$$\frac{\partial p}{\partial y \partial t}=0 \qquad \text{(alternative 2)}$$

Suppose now that alternative (2) holds. It follows then from eqs.(32) and (34) that

$$\boxed{\frac{\partial u}{\partial t}=0} \tag{38}$$

Eq.(25) then gives

$$\boxed{\frac{\partial u}{\partial x}=0} \tag{39}$$

and eq.(29) reduces to

$$\boxed{\frac{\partial \rho_s}{\partial x}=0} \tag{40}$$

Furthermore, from eq.(28)

$$\boxed{\frac{\partial v}{\partial z}=0} \tag{41}$$

Conversely, if we adopt alternative (1) eqs.(29),

(25) and (32) show that

$$\boxed{\frac{\partial^2 p}{\partial y \partial t}=0} \tag{42}$$

Thus, both the alternatives (1) and (2) must be satisfied in general geostrophic motion, i.e. the isopycnics must coincide with the isobars, and there must be no isallobaric gradient.

13. Vorticity.

Let q_x, q_y and q_z denote the components of the vorticity vector. In accelerated motion*

$$\frac{dq_z}{dt}=-\left(\frac{\partial \alpha}{\partial x}\cdot\frac{\partial p}{\partial y}-\frac{\partial \alpha}{\partial y}\cdot\frac{\partial p}{\partial x}\right)+\left(q_x+2\omega_x\right)\frac{\partial w}{\partial x}+\left(q_y+2\omega_y\right)\frac{\partial w}{\partial y}$$
$$-\left(q_z+2\omega_z\right)\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)$$

where $\omega_x, \omega_y, \omega_z$ are the components of the earth's rotation at the point considered.

The conditions for general geostrophic motion substituted into the vorticity equation gives :

$$\boxed{\frac{dq_z}{dt}=0} \tag{43}$$

Thus, the vorticity in the horizontal plane is conserved, and it can be shown similarly that the same is true also of the other components.

The above theorem does not hold for singular geostrophic motion. In this case

$$-\left(\frac{\partial \alpha}{\partial x}\cdot\frac{\partial p}{\partial y}-\frac{\partial \alpha}{\partial y}\cdot\frac{\partial p}{\partial x}\right)=2\omega\sin\varphi\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)$$

and the vorticity equation reduces to

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*See for example Sverre Petterssen and James M.Austin:"Fronts and Frontogenisis in relation to Vorticity", Papers in Physical Oceanography and Meteorology, Vol.VII, No.2, 1942.

[redacted]

$$\frac{dq_z}{dt} = -\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) q_z$$

14. The First Law of Thermodynamics.

If Q, T and C_v denote heat, absolute temperatures, and specific heat at constant volume, the first law of thermodynamics may be written :

$$\frac{dQ}{dt} = C_v \frac{dT}{dt} + p\frac{d\alpha}{dt}$$

Since, when the x-axis is along the isobar

$$\frac{\partial \rho_s}{\partial t} = \frac{\partial \rho_s}{\partial x} = 0 \quad \text{and} \quad V = W = 0$$

in geostrophic motion, it follows that

$$\frac{d\rho_s}{dt} = \frac{d\alpha}{dt} = 0 \qquad (44)$$

Thus, expansion is incompatible with geostrophic motion, and therefore,

$$\frac{dT}{dt} = \frac{dp}{dt} = 0 \qquad (45)$$

Since the motion is horizontal and along the isobars

$$\frac{dp}{dt} = \frac{\partial p}{\partial t}$$

and therefore from eq.(45)

$$\boxed{\frac{\partial p}{\partial t} = 0} \qquad (46)$$

Thus, in general geostrophic motion, the pressure is stationary.

15. Criteria for Geostrophic Motion.

The results of the discussion in the foregoing paragraphs may now be summarized as follows:

Motion that is geostrophic not only at a singular point (x,y, z) but also within a small volume of air around that point, possesses the following characteristics.*

1. The isobars in a horizontal plane must be straight and parallel (22, 23).
2. The isobars must not rotate with increasing or decreasing elevation (40 and 28).
3. The barometric tendency must be zero (46, 27, 36, 42).
4. The wind is stationary and constant along each isobar (26, 38, 39)
5. The wind direction must not vary with elevation (41).
6. The divergence is zero (26, 39, 44).
7. The vorticity is conserved (43).
8. The density is stationary and so is also the temperature (35,46)
9. The isopycnics, and therefore, the isosterics and the isotherms must be parallel to the isobars in level surfaces (40).

The question now arises; do the above characteristics represent nine mutually independent conditions that must be satisfied simultaneously or are they inter-related in such a manner that they can be reduced to one, or more, fundamental conditions. An analysis of the derivation of the above characteristics will show that they can all be reduced to three fundamental criteria that must be satisfied for a volume of air to move geostrophically. These criteria are :

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*Numbers in parentheses identify the equations to which the statements refer.

	For isobaric charts at constant levels.	For pressure contour charts.
A	The isobars must be straight and parallel.	The contours must be straight and parallel
B	The isopycnics (and isosterics and isotherms) must be parallel to the isobars.	The thickness lines must be parallel to the contours.
C	The barometric tendency must be zero.	

From criterion B, and also from the characteristics 5 and 9 above, it follows that the geostrophic wind cannot transport thickness lines.

There is, therefore, no thermal contribution to the pressure changes, and criterion C then states that the barometric tendency must be zero at the base of the layer that moves geostrophically. Thus, for example, if the barometric tendency at sea level (where observations are copious) is very small and if the contours of (say) the 750 mb. surface are almost straight and parallel to the sea level isobars (i.e. thickness lines parallel to the contours) then the motion is for all practical purposes, geostrophic. On the other hand, if the barometric tendency at sea level is appreciable, and/or if the contours of (say) the 750 mb. surface form an angle with the isobars at sea level, then the motion is not geostrophic, and it is unbalanced in proportion to the existing deviations.

Thus, the criteria for the existence of geostrophic balance are such that their fulfilment, or non-fulfilment, can readily be ascertained by the aid of synoptic charts.

16. Vertical Variation of the Geostrophic Wind.

Returning now to eq.(31) and introducing specific volume instead of density we obtain:

$$\frac{\partial u}{\partial z} = \frac{1}{\alpha}\left[\frac{\partial \alpha}{\partial z}u - \frac{g}{2\omega\sin\varphi}\cdot\frac{\partial \alpha}{\partial y}\right]$$

Let B be the slope of the pressure surface and C the slope of the x-surface, then when the x-axis is chosen along the isobar

$$\frac{\partial \alpha}{\partial x} = 0$$

and

$$\tan C = -\frac{\partial \alpha}{\partial y} \Big/ \frac{\partial \alpha}{\partial z}$$

Substituting this in the above equation

$$\frac{\partial u}{\partial z} = \frac{1}{\alpha}\frac{\partial \alpha}{\partial z}\frac{g}{2\omega\sin\varphi}\left(\tan C - \tan B\right) \qquad (47)$$

Simultaneously eq.(41) holds. Therefore, in general geostrophic motion the wind direction does not vary with height,* and the speed of the wind increases along the vertical when the slope of the isosteric surface exceeds that of the isobaric surface. If the isosteric surfaces coincide with the isobaric surfaces, the stratification is barotropic; the wind (and the Coriolis force) is then constant along the vertical.

A convenient expression for the variation in geostrophic wind through a finite layer of air is readily obtained from eq.(18), when the x-axis is chosen along a contour line. Writing down the equations for each of the pressure surfaces (1) and (2) and subtracting, we obtain:

$$\Delta V = 0$$

$$\Delta u = \frac{-g}{2\omega\sin\varphi}\left(\tan B_2 - \tan B_1\right) \qquad (48)$$

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*This condition is expressed by criterion B in para 15 and eq.(41).

Thus, the increase in speed from a pressure surface p_1 to a pressure surface p_2 is proportional to the relative slope of the two surfaces, or to the horizontal gradient of thickness lines.

17. Three Concepts of Geostrophic Motion.

In view of the fact that many of the above results are at varience with those appearing in current literature, it may be of interest to comment upon the sources of the discrepancies.

(1) The singular type. The geostrophic motion discussed in most texts is of the singular type (see para.7), for the only restriction imposed is that

$$\dot{u} = \dot{v} = 0$$

Usually, it is not even demanded that the motion should satisfy the equation of continuity, and, without this requirement, erratic theoretical results may be obtained.

In practice, the singular type of geostrophic wind is used as an approximation to the true wind, and as such it has rendered the most excellent services. The approximation is obtained by neglecting the accelerations in individual points, and thus equating the pressure force unto the Coriolis force, without ascertaining that balanced motion is dynamically possible under the existing conditions.

Obviously as a practical approximation this kind of geostrophic wind can possess no particular dynamical properties of its own: it must possess approximately the divergence, the curl, the accelerations (local and individual) of the wind that it is intended to approximate. If the density varies along the isobars, the approximated wind must transport density (and thickness lines) and be held responsible for developments; and if the isobars rotate horizontally, or vertically, the approximated wind must do likewise. If the actual wind is geostrophic, the approximation is exact, and the more unbalanced the motion the poorer the approximation.

(2) Expression for the Pressure Force. In theoretical discussions it is sometimes convenient to substitute an entirely artificial "geostrophic" wind for the pressure force. Thus, if G_x and G_y are the components of a wind such that the Coriolis force corresponding thereto would balance the existing pressure force (whether or not the isobars are curved, moving etc.) the equations of horizontal motion could be written:

$$\left.\begin{aligned}\dot{u} &= 2\omega \sin\varphi\,(v - G_y)\\ \dot{v} &= -2\omega \sin\varphi\,(u - G_x)\end{aligned}\right\} \qquad (49)$$

It is evident that this artificial wind is dynamically unrestricted: it represents an identical expression of the pressure force, and considered as a wind it has no reality. The damage that this artificial wind does (e.g. by appearing to cause a transport of thickness lines, etc.) must be repaired by an equally artificial wind (i.e. the geostrophic departure) so as to render the results that are caused solely by the actual wind. There can, of course, be no harm in introducing an artificial wind to represent the pressure force as long as the results are not interpreted as representing the action of the wind that results from balance of the forces.

(3) Dynamic Reality. The geostrophic motion whose properties have been elaborated above, represent a dynamical reality, namely the motion that results when a volume of air (which may be large or small) moves under balanced forces. It satisfies not only the equations of unaccelerated motion but also the equation of continuity, the hydrostatic equation, etc., in it, no work is done by any of the forces, there is no divergence and no development. The criteria for this balance to occur are such that their fulfilment, or non-fulfilment, can readily be

ascertained in each individual case. The application of these criteria to synoptic charts will enable the synoptic Meteorologist to identify balanced and unbalanced states.

To illustrate further the difference between this geostrophic motion and the singular type, we consider the distribution of $\dot{u}$ along the x-axis at a given instant, as shown in Fig.2.

Curve I represents the singular case, that is when the acceleration vanishes only at a point x_0. Curve II, on the other hand, represents a case in which the acceleration vanishes not only at the point x_0, but, with high degree of accuracy, also within the internal $x_0 - \Delta x$ to $x_0 + \Delta x$. Thus, in the case represented by Curve I

$$\frac{\partial \dot{u}}{\partial x} \gtrless 0$$

and in the case represented by Curve II

$$\frac{\partial \dot{u}}{\partial x} = 0$$

The criteria for general geostrophic motion result through applying this latter criterion also to y and z.

18. The Thermal Wind.

Returning now to eqs. (19) and integrating from z_0 to z_1 we obtain

$$-\left[\left(\frac{\partial p}{\partial x}\right)_1 - \left(\frac{\partial p}{\partial x}\right)_0\right] = g\int_{z_0}^{z_1} \frac{\partial \rho_s}{\partial x}\,dz = g\,\frac{\partial \rho_m}{\partial x}$$
$$-\left[\left(\frac{\partial p}{\partial y}\right)_1 - \left(\frac{\partial p}{\partial y}\right)_0\right] = g\int_{z_0}^{z_1} \frac{\partial \rho_s}{\partial y}\,dz = g\,\frac{\partial \rho_m}{\partial y}$$

Here, ρ_m is the mean density of the air column from z_0 to z_1.

It should be noted that the above equations are derived from the hydrostatic equation; they hold, therefore, in accelerated as well as in balanced motion. In vector rotations, the equations state that the pressure gradient at the level z_1 is made up of two parts, viz., the pressure gradient at the level z_0 plus a pressure gradient that is equal to the acceleration of gravity times the horizontal ascendant of the mean density of the air column as shown in Fig. 3.

If we now resort to the second concept of the geostrophic wind (i.e. (2) in para. 17) and substitute the artificial geostrophic wind as a mere expression for the pressure force, we obtain

$$\left.\begin{aligned} V_1 &= \frac{\rho_0}{\rho_1} V_0 - \frac{g}{2\omega \sin\varphi}\frac{\partial \rho_m}{\partial x} = \frac{\rho_0}{\rho_1} V_0 - V_T \\ u_1 &= \frac{\rho_0}{\rho_1} u_0 + \frac{g}{2\omega \sin\varphi}\frac{\partial \rho_m}{\partial y} = \frac{\rho_0}{\rho_1} V_0 + U_T \end{aligned}\right\} \qquad (50)$$

where U_T and V_T are the components of what is usually called the thermal wind*. Thus, if we introduce the artificial geostrophic wind as a mere expression for the pressure force, we may say that the geostrophic wind at the level z_1 is equal to the geostrophic wind at the level z_0 multiplied by the density ratio ρ_0/ρ_1 + the thermal wind, which is proportional to, at right angles to and to the right of the ascendant of the mean density of the air column from z_0 to z_1. Thus, the thermal wind blows along the mean density lines with higher density 'and lower temperature) to its left. This statement holds, of course, in general motion and must therefore contain geostrophic balance as a special case. As was shown in para.16, geostrophic balance requires that the density is constant along each isobar. With the x-axis along the isobar at the level z_0.

$$V_0 = 0 \qquad\qquad \frac{\partial \rho_m}{\partial x} = 0$$

and eqs.(50) reduce to eqs.(48).

*In discussions of the wind variation with height it is customary to [illegible] temperature instead of density; hence the expression "thermal wind"

Next, we may resort to the first concept of the geostrophic wind, and interpret U and V in eqs.(50) as geostrophic approximations to the true wind. All we do then is to neglect the accelerations at the base and at the top of the air column without assuming that the air in between the levels z_0 and z_1 is under geostrophic balance. This is the same as saying that the winds at each of the two levels z_0 and z_1 are approximately geostrophic, but nothing is implied with regard to the distribution of the accelerations within the air column.

The above derivation, although mathematically inelegant, is reproduced here in order to underline the importance of the density distribution and remove certain misunderstandings concerning the true nature of the so-called thermal wind. A more elegant derivation is obtained by substituting for α from the hydrostatic equation into the equations of motion. We then obtain for the levels z_0 and z_1 respectively:

$$\dot{u}_1 = -g \tan A_1 + 2\omega \sin\varphi\, v_1 \qquad \dot{v}_1 = -g \tan B_1 - 2\omega \sin\varphi\, u_1$$
$$\dot{u}_0 = -g \tan A_0 + 2\omega \sin\varphi\, v_0 \qquad \dot{v}_0 = -g \tan B_0 - 2\omega \sin\varphi\, u_0$$

To obtain the so-called thermal wind we need not assume that there is approximate (or exact) geostrophic balance at any level. It suffices to assume that

$$\dot{u}_1 \approx \dot{u}_0 \qquad \text{and} \qquad \dot{v}_1 \approx \dot{v}_0$$

and the accelerations may have any value between the levels considered. Subtraction then gives:

$$\begin{aligned} v_1 - v_0 &= \frac{g}{2\omega \sin\varphi}\left(\tan A_1 - \tan A_0\right) \\ u_1 - u_0 &= -\frac{g}{2\omega \sin\varphi}\left(\tan B_1 - \tan B_0\right) \end{aligned} \qquad (51)$$

These equations* hold in accelerated motion, with the slight restriction that the accelerations at the base of the layer must be approximately equal to those at the top of the layer; they can, therefore, not be taken to characterise geostrophic balance. If geophstrophic balance obtains within the layer, eqs.(51) reduce to eqs.(48), which represent real geostrophic balance.

IV ACCELERATIONS.

19. The Synoptic Problem.

The results derived in the foregoing paragraphs indicate that it is possible to ascertain from analysed upper air charts the presence, or absence, of balance in atmospheric motion. In other words, it has now become possible to improve upon the present practice, which consists in applying the geostrophic wind scale under all circumstances, which, again, amounts to demanding that the wind shall be geostrophic in all possible combinations of distributions of mass versus pressure.

Considering the acceleration problem from a synoptic point of view and taking stock of the situation, two facts become apparent, viz.,

(1) That one possesses no experience in mapping and analysing the accelerational field (the need for experimentation and development of a technique of analysis is, therefore obvious), and

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* In these equations the density does not enter explicitly, but the difference in slope of the pressure surfaces through z_0 and z_1 is uniquely determined by the distribution of density.

(2) That it is known that the accelerations may have three different and rather distinct missions, namely:

(a) To rotate the winds without appreciably creating kinetic energy, (as in the case in gradient wind balance where the accelerations are entirely, or mainly, centripetal).

(b) To cause the pressure systems to travel, and

(c) To cause developments in the pressure field. Thus, for example, even if it were known that the acceleration in a certain place was, say, 2.6×10^{-3}, the forecaster would hardly know what conclusions to draw unless he knew whether this was a large, medium, or small value, and was able to refer the effect of the acceleration to either of the above categories (a), (b) and (c).

Having in the foregoing paragraphs developed criteria by the aid of which unbalanced motion can be discovered, the remaining problem appears to be to develop methods by the aid of which the accelerations (or the significant effects of the accelerations) can be assessed and classified. A logical approach towards the solution of this problem would be to investigate a large number of situations, including stationary and travelling, deepening and filling cyclones, anti-cyclones etc. particularly with a view to:

(i) Developing a technique of dynamical analysis
(ii) Gaining further experience, and
(iii) Determining the practical usefulness of the various criteria.

In synoptic meteorology a synthesized assessment of the factors involved is always more fruitful that a strict numerical evaluation of individual elements at single points, and the method of assessing accelerations (or similar quantities) can be developed only through experimentation. Until such experiments can be made, it appears unnecessary here to enter into detailed theoretical discussion of the acceleration problem, particularly since the very formulation of the problems would be largely influenced by the results of the practical investigations. The following brief remarks on accelerations are, therefore, intended only to indicate some of the possibilities that now appear to exist.

The most obvious approach would be to evaluate the accelerations from actual observations and segregate the accelerational components mentioned under (a), (b) and (c) above. This method, although inelegant in theory and cumbersome in practice, need not be altogether void of practical possibilities. Some brief comments on this question will be found in para. 21.

A far more promising approach, however, appears to reside in the dynamical interpretation of the mass pressure patterns, as revealed by the contours and thickness lines of isobaric sheets. What the forecasters need more than explicit expressions for the accelerations are separate expressions (or criteria) for the speed and the development of pressure systems in which the accelerations do not enter explicitly. Experience gained by analysing upper air charts seems to suggest that a formulation of the problem along these lines offer the best chances for practical results.

20. Mass-Pressure Patterns.

We shall now consider an isobaric layer of air determined by the limiting pressures

P_1 and P_2 $(P_1 > P_2)$

Suppose now that the contours of these two pressure surfaces are drawn on one chart, as indicated diagramatically in Fig.4. In general, the two sets of contours will intersect forming a grid of quadrilaterals; and, ceteris paribus, the deeper the layer the larger the angle between the two sets of contours.

Now if the motion were geostrophic within the layer, the two sets of contours would be parallel. On the other hand, the more unbalanced the motion the more pronounced is the grid, and for sufficiently deep layers (e.g. $P_1 - P_2 = 200$ mb., or more) the angle between the contours will (if there is any appreciable acceleration) be large in comparison to the angle that may result on account of errors in analysis and in perfection of instruments. Thus, through considering layers of sufficient depth, significant mean values can be obtained.

The two sets of contours (i.e. H_1 and H_2 in Fig.4) indicate more than the mere geometry of the pressure surfaces. Thus, the diagonal D (i.e. the difference diagonal) indicates the thickness of the layer, and sets of such diagonals show the distribution of the thickness of the isobaric sheet. These diagonals will hereafter be referred to as thickness lines. The dynamical significance of the thickness lines resides in the fact that the mean specific volume (and the mean temperature) of the layer is constant along each thickness line.

The other diagonal M (i.e. the sum diagonal) represents the mean contour of the layer, and the set of such diagonals represents the mean pressure distribution within the layer.

Directing now our attention to the sets of diagonals rather than to the contours themselves, we have a fair picture of the horizontal mass-pressure pattern of the layer. The grid formed by these diagonals represents the isosteric-isobaric solenoids within the layer. Through these diagonals the chart becomes dynamically alive; in balanced motion the grid of diagonals disappears, and the grid that exists in any given case is expressive of the distribution of the unbalanced part of the motion.

The mass-pressure patterns appear most distinctly, in shape and in intensity, in connection with fronts and cyclones, and here we meet with a variety of types that deserve closer examination. Specific types will not be discussed in this paper, but in order to illustrate the essence of these patterns we shall consider a schematical cross section through a warm sector cyclone as shown in Fig.5.

As an example, we consider the height of the 700 mb. surface above the 1000 mb. surface. The column CC' is entirely within the cold air, and the column WW' is entirely within the warm air. The thickness from the 1000 to the 700 mb. surface must therefore decrease as we proceed in the direction from W to C, and the slope of the pressure surfaces increase with increasing elevation. If now a chart is drawn for each of the pressure levels, it will be seen that on each of these charts a frontal region appears with a concentration of thickness lines within the belt between the position of the front at sea level and the position of the front at the level considered. The width of this frontal region increases with increasing elevation. Furthermore, for any given overall temperature difference between the air masses involved the gradient of thickness increases in proportion to the slope of the front.

It will be seen from Fig.5 that although the two air masses may be regarded as being separated by a frontal surface, the dynamically significant frontal region consists of a volume of air, resting upon the cold wedge and having a more or less triangular cross section. It is within this region that there is a maximum of concentration of thickness lines. Thus, for example, the frontal region below the 700 mb. surface in Fig.5 has the cross section indicated by the triangle WW'C'.

The total topography of either of the pressure surfaces in Fig.5 is obtained by adding graphically the diagram C to the diagram B. It will then be seen that, depending on the overall temperature difference, the slope of the fronts and the topography of the 1000 mb. surface, the downward slope within the frontal region in diagram C may, or may not, be compensated for by the upward slope of the 1000 mb. surface in diagram B. But since the downward slope of the surfaces in diagram C increases with increasing elevation, there will, in general, be a level at which compensation is established. Although little has been done to explore these conditions, it can be said that the level of compensation is usually well below 500 mb. level, although in some cases it has been found to reach to higher levels.

Below the level of compensation, the isobars (and the contours) within the frontal region will have a cyclonic trend, and above that level the trend will be anti-cyclonic. Thus, for example, if the contours of the 900 and 600 mb. surfaces corresponding to Fig.5 are drawn, the diagrams shown in Figs. 6 and 7 result.

In order not to overload the diagrams with curves, the diagonal grids have been omitted. It will, however, be seen that the mass-pressure patterns established in these models are such that there is a tendency for a maximum of thermal wind[*] to occur above the pressure centre at low levels. Experience, which needs further corroboration, seems to indicate that, ceteris paribus, the speed of the pressure system is proportional to this thermal wind; that is, to the gradient of thickness over the pressure centre perpendicular to the paths. Thus, if the x-axis is tangent to the front in the pressure centre the speed of the centre appears to be proportional to

$$\int_{z_0}^{z_1} \frac{\partial \rho_s}{\partial y} dz$$

The practical problem here appears to be to determine the limits of this integral that, under any given conditions, will render the highest correlation with the observed speeds. This might, eventually, lead to a method of assessing the part of the unbalanced motion that determines the speed of a pressure system.

The above rule appears in particular to afford a mass-pressure criterion for stationarity of pressure systems. In a few cases examined it was invariably found that stationary cyclones and anticyclones were characterized by coincidence of contours and thickness lines, suggesting either no acceleration, or purely centripetal acceleration.

Figs. 5 to 7 represent mere diagrams and suffer from several over-simplifications, including lateral symmetry. In "Weather Analysis and Forecasting" attention was drawn to certain types of wave cyclones (as they appear on surface charts) which experience has shown to deepen rapidly. These waves were all characterized by lateral asymmetry in the pressure distribution. Recent experience seems to indicate that asymmetry in the mass-pressure pattern is an important factor to consider, but, so far, no satisfactory criterion to distinguish deepening from filling can be suggested. In fact, it appears that the distinction between deepening and filling resides in the degree rather than in the type of asymmetry.

The writer desires to emphasize that the above discussion of the influence of the mass-pressure patterns on the behaviour of cyclones should be taken as suggestions of problems for further research, and not as established facts. He believes, however, that, through further studies, it will be possible to derive expressions for the speed and the development of pressure systems in which the accelerations are accounted for implicitly. In[x] this connection mention should be made of a recent work by Rossby in which he discusses frontless pressure waves in the Prevailing

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[*] The term "Thermal wind" is here used in the general meaning indicated in para. 1B.

[x] For footnote see page 19.

[redacted]

Westerlies. From purely kinematical consideration based upon largely restricting assumptions, he obtains some highly interesting results, showing that there are certain pattern-criteria for the speed and the increase in amplitude of the waves. In these expressions the accelerations do not enter explicitly.

21. Mean Accelerations.

Under favourable circumstances, it is entirely possible to evaluate the mean acceleration and the vertical shear of acceleration within isobaric sheets. As in the foregoing paragraph we consider an isobaric sheet determined by the limiting pressures p_1 and p_2 (see Fig.4). Along an arbitrary x-axis:

$$\dot{u} = -g \tan A + 2\omega \sin\varphi \, v$$

Writing this equation for each of the pressure levels, adding and subtracting, we obtain

$$\dot{u}_m = -g \frac{\tan A_2 + \tan A_1}{2} + 2\omega \sin\varphi \, v_m \qquad (52)$$

$$\Delta\dot{u}_m = -g(\tan A_2 - \tan A_1) + 2\omega \sin\varphi \, \Delta v \qquad (53)$$

where subscript $_m$ indicates the mean value, and Δ the increase through the layer.

In general, these equations possess no advantage over the ordinary equations of motion, for the accelerations appear as the difference between the Coriolis force and the pressure force. However, through suitable choice of system of co-ordinates, certain simplifications result.

It will be seen from Fig.4 that if we proceed along the difference diagonal D in the direction from a to b, we will ascend the same height along both pressure surfaces. Therefore, if we choose the x-axis along the thickness line

$$\tan A_1 = \tan A_2$$

and the above equations reduce to

$$\dot{u}_m = -g \tan A + 2\omega \sin\varphi \, v_m \qquad (54)$$

$$\Delta\dot{u}_m = 2\omega \sin\varphi \, \Delta v \qquad (55)$$

The last of these equations states that the increase in acceleration along the thickness line from the base to the top of the layer is proportional to the increase in wind normal to the thickness lines.

Now, if geostrophic balance obtains, the thickness lines are parallel to the contours, and,

$$v = \Delta v = 0$$

and the mass-pressure pattern is stationary.

If, on the other hand,

$$v \gtrless 0 \qquad \text{and} \qquad \Delta v = 0$$

the mass-pressure pattern will be non-stationary, but there will be no deformation of the pattern from one layer to another. This

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*C-G Rossby: Kinematic and Hydrostatic Properties of certain Long Waves in the Prevailing Westerlies. The University of Chicago, Inst. of Met. Miscellaneous Reports, No.5, 1942.

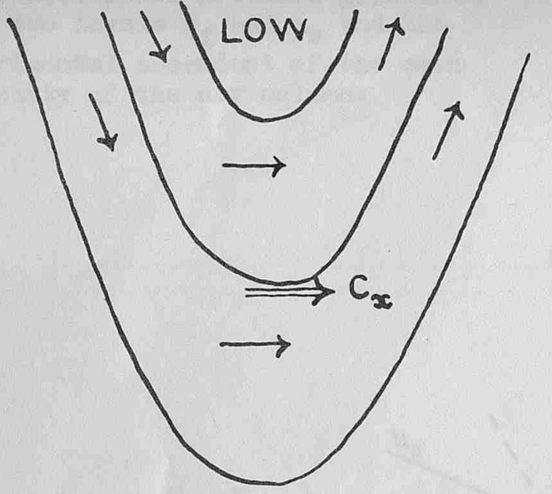


Fig.1

→ Wind vectors

⇒ speed of pressure trough

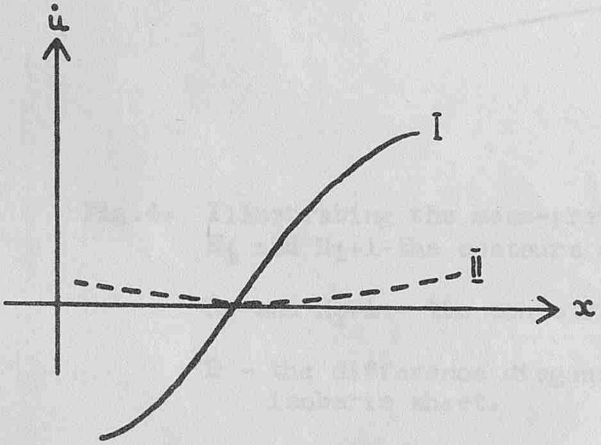
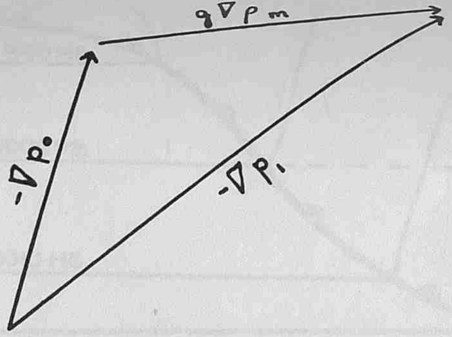


Fig. 2.

Curve I represents a case in which $\dot{u}$ vanishes in a singular point only.

Curve II represents a case in which $\dot{u}$ may be regarded as vanishing within an interval.

Fig.3.
Showing the relation between **the horizontal** pressure gradients at two levels p_1 and p_0 and the horizontal ascendant of the mean density of the air column.



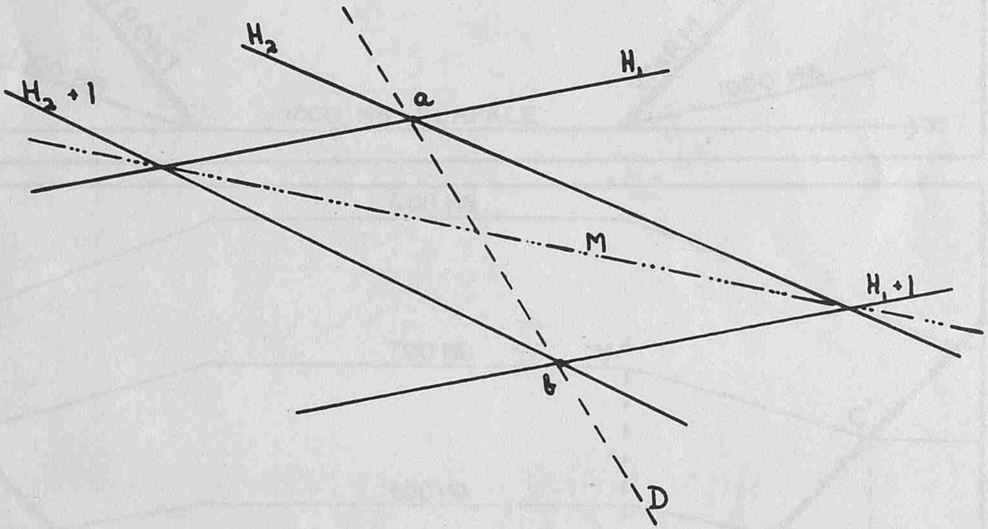


Fig.4. Illustrating the mass-pressure distribution within a layer H_1 and H_1+1-the contours of the lower pressure surface.

H_2 and H_2+1 - the contours of the upper pressure surface.

D - the difference diagonal, or the thickness of the isobaric sheet.

M - the sum diagonal, or the mean contour of the isobaric sheet.

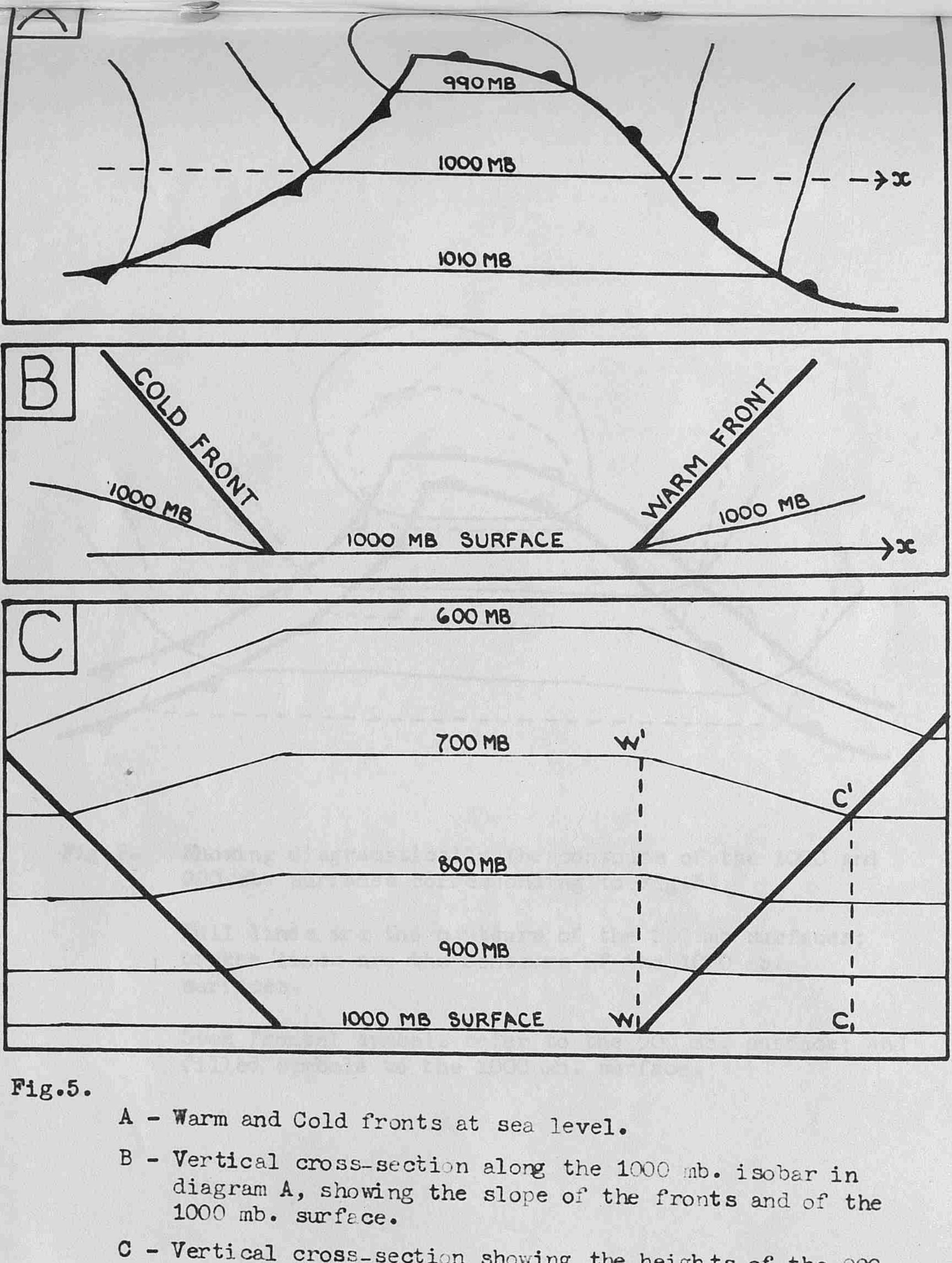

A
990 MB
1000 MB
1010 MB
x
B
COLD FRONT
WARM FRONT
1000 MB
1000 MB
1000 MB SURFACE
x
C
600 MB
700 MB
W'
C'
800 MB
900 MB
1000 MB SURFACE
W
C

Fig.5.

A - Warm and Cold fronts at sea level.

B - Vertical cross-section along the 1000 mb. isobar in diagram A, showing the slope of the fronts and of the 1000 mb. surface.

C - Vertical cross-section showing the heights of the 900, 800, 700, and 600 mb. surfaces above the 1000 mb. surface.

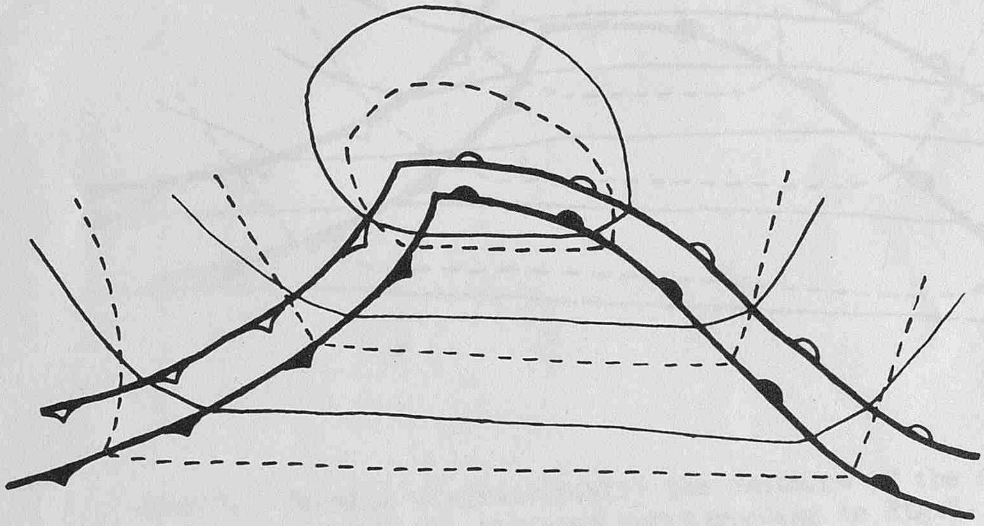


Fig.6. Showing diagramatically the contours of the 1000 and 900 mb. surfaces corresponding to Fig.5.

Full lines are the contours of the 900 mb surfaces; broken lines are the contours of the 1000 mb. surfaces.

Open frontal symbols refer to the 900 mb. surface; and filled symbols to the 1000 mb. surface.

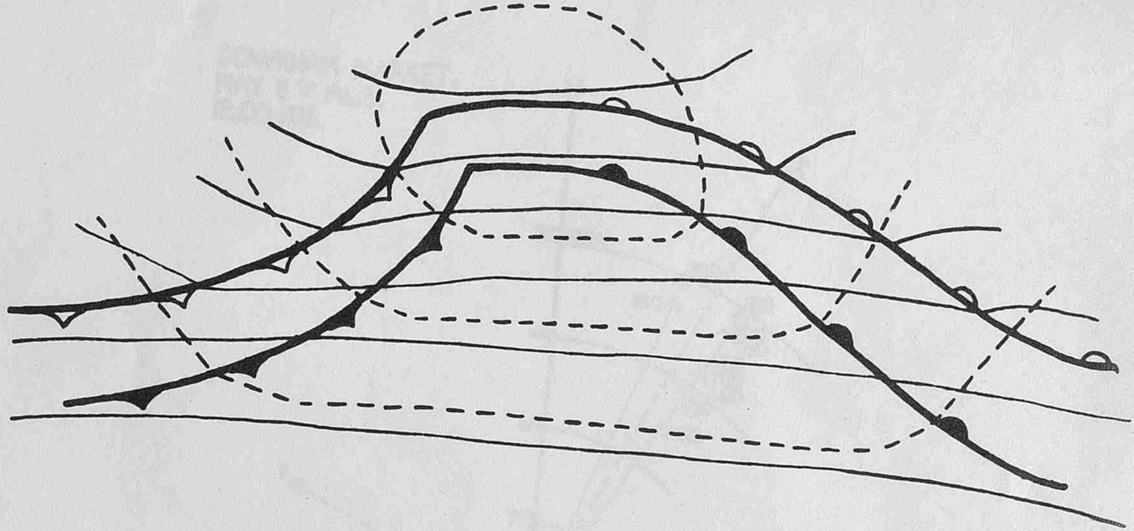


Fig.7. Showing diagramatically the contours of the 900 mb. and the 600 mb. surfaces corresponding to Fig.5.

Full lines are the contours of the 600 mb. surface; broken lines are the contours of the 900 mb. surface.

Open frontal symbols refer to the 600 mb. surface, and filled symbols to the 900 mb. surface.

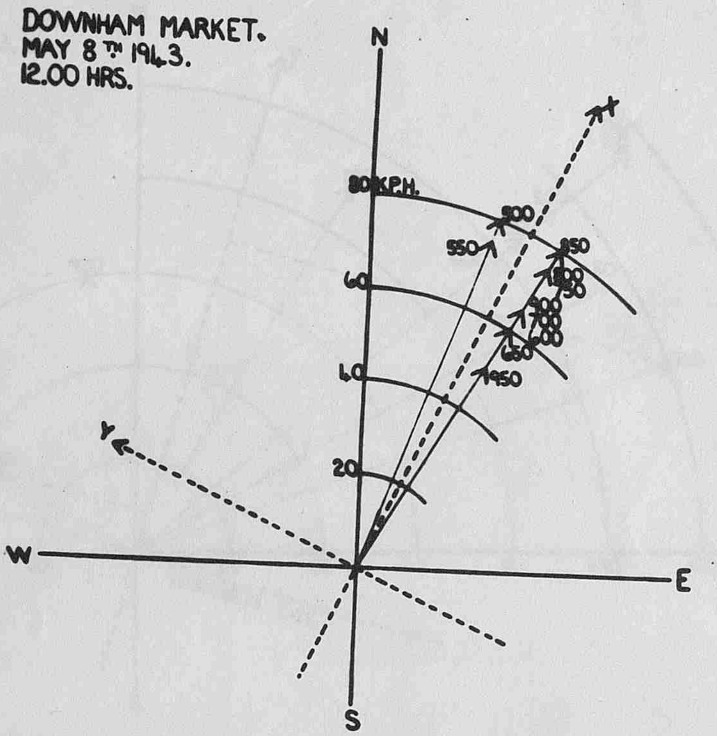


Fig. 8.- Wind variation along the vertical suggesting geostrophic balance. Isobars and thickness lines are parallel to the winds.

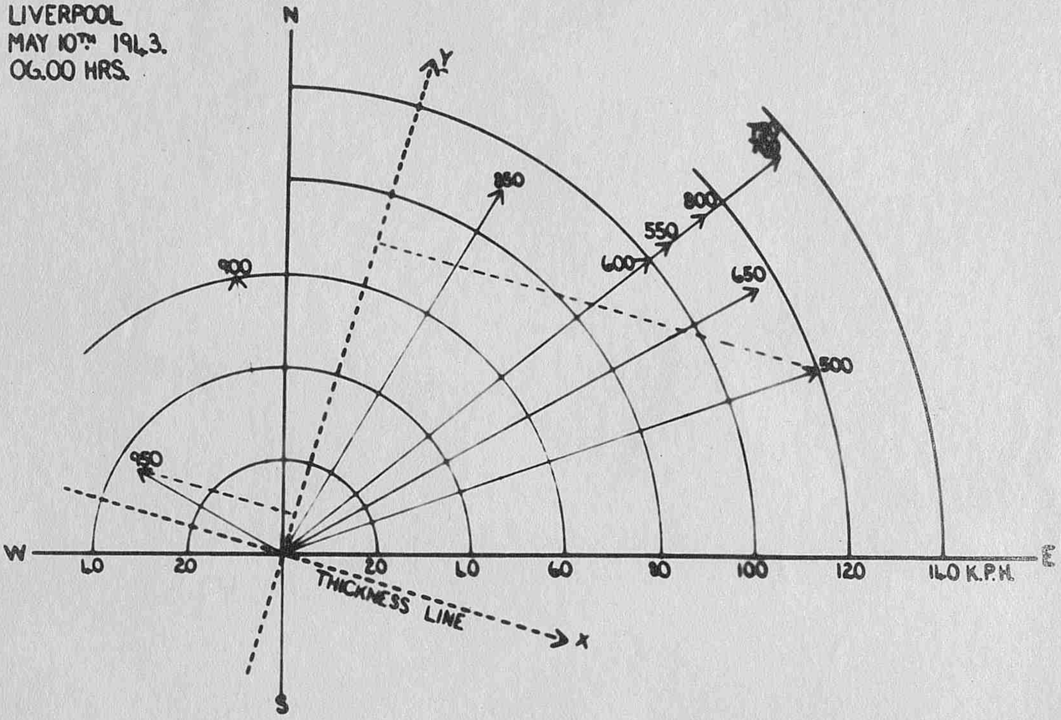


Fig. ~~13~~ 9.- Wind variation along the vertical showing unbalanced motion. The x-axis is chosen along the thickness line of the layer from ~~1,000~~ 950 to 500 mb.

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Mention Cond. Trails pp. 2-3

on p. 2 of classif. log -- RTF

AIR MINISTRY

M.R.P.150
30 October 1943

METEOROLOGICAL RESEARCH COMMITTEE

Maximum and Minimum Temperatures in the Upper Air.

by

F.M.Jones.

AIR WEATHER SERVICE
TECHNICAL REFERENCE FILE

Questions sometimes arise regarding the highest or lowest temperatures likely to occur at various heights. Figures 1 and 2 show the highest and lowest recorded in July and August, and in January and February respectively in the following cases.-

(a) In England - ballons sondes from 1921 - 37.[1,2]

(b) In Alaska - observations from 7 stations, the records being from 2 to 4 years duration.

(c) In U.S.A.(Northern States) - observations from 28 stations, the records being from 1 to 11 years duration.

(d) In U.S.A.(Southern States) - observations from 20 stations, from 1½ to 8½ years duration.

The American tabulations [3,4] are based on aircraft, radio sonde and kite observations. The kite observations were replaced by aircraft observations during 1931 - 33. Radio sonde observations were started in 1938.

The United States were, for the purpose of this analysis, divided into two groups, as follows.-

(i) Northern States.- California, Colorado, Dakota, Idaho, Illinois, Massachusetts, Maine, Michigan, Minnesota, Montana, Nebraska, Nevada, New Jersey, New York, Ohio, Oregon, Utah, Washington and Wyoming.

(ii) Southern States.- Alabama, Arizona, Canal Zone, Carolina, Florida, Georgia, Louisiana, Missouri, New Mexico, Oklahoma, Tennesse, Texas and Virginia.

Maximum Temperatures.

Troposphere. In summer and also in winter the highest temperatures attained in England are not much different from those in Alaska. The Northern and Southern States approximate to one another, both experiencing maximum temperatures some 10 to 15°C higher than England or Alaska.

Stratosphere. In summer above 12 km. the Northern States and Alaska get the highest maxima, and the Southern States the lowest maxima. In winter Alaska above 11 km. gets the highest temperatures, and Engalnd curiously enough gets the least high. It will be noted that the maximum stratosphere temperatures at Alaska are much the same in winter as in summer.

Minimum Temperatures.

Troposphere. In summer, from 1 to about 11 km., conditions of minimum temperature in the Northern States approximate to those in England. In Alaska it is liable to be from 3 to 10°C colder. In the Southern States the limits are some 3 to 8°C higher. In winter the conditions in England come nearer to those of the Southern States. In the Northern States, above 5 km., conditions tend to be the same as in Alaska.

Stratosphere. Both in summer and in winter, conditions of minimum temperature in Alaska and England are closely similar above 12 km.

Much lower temperatures are attainable in both Northern and Southern States, and particularly in the Southern States above 14 km. The minimum temperatures above 14 km. in the Southern States are in effect much the same as the temperature of the stratosphere in the tropics, and they vary very little from summer to winter. The fact that these low temperatures are not found in higher latitudes than the Southern States suggests either that the air of the tropical stratosphere does not move to higher latitudes, or that, if it does move to higher latitudes, it is only under conditions which change its temperature appreciably.

Exhaust Condensation Trails.

The immunity line for condensation trails has been entered in figures 1 and 2 as II.

It looks as if on occasion conditions could be cold enough to allow of the formation of condensation trails at any level in winter, in Alaska and the Northern States; while in summer, even under the coldest conditions, such trails should not occur below 6 km. in any of the four regions considered.

Table 1a.- Maximum Temperatures in July and August.

Height km.	Alaska °C	Northern States °C	Southern States °C	England °C
1	24	37	40	22
2	17	29	33	12
3	11	23	24	9
4	5	14	14	2
5	0	8	7	-4
6	-6	3	0	-10
7	-14	-5	-5	-16
8	-22	-13	-10	-25
9	-30	-20	-16	-30
10	-36	-25	-22	-38
11	-38	-31	-30	-38
12	-39	-37	-37	-42
13	-38	-37	-43	-39
14	-39	-39	-50	-44
15	-39	-40	-54	-41
16	-40	-40	-56	-44

Table 1b.- Minimum Temperatures in July and August.

Height km	Alaska °C	Northern States °C	Southern States °C	England °C
1	-6	4	11	5
2	-10	-5	3	-2
3	-14	-8	-1	-6
4	-19	-14	-8	-12
5	-24	-21	-11	-19
6	-32	-29	-20	-26
7	-40	-36	-27	-34
8	-45	-43	-34	-42
9	-52	-51	-43	-48
10	-57	-54	-48	-53
11	-61	-59	-57	-58
12	-63	-67	-67	-62
13	-63	-75	-72	-64
14	-62	-74	-77	-61
15	-59	-80	-84	-62
16	-59	-77	-89	-60

Table 2a.- Maximum Temperatures in January and February.

Height km	Alaska °C	Northern States °C	Southern States °C	England °C
1	9	23	26	10
2	9	16	23	3
3	1	12	17	-1
4	-7	5	9	-6
5	-13	-2	4	-13
6	-19	-8	-3	-20
7	-25	-14	-10	-26
8	-34	-22	-17	-33
9	-39	-29	-25	-44
10	-39	-35	-31	-43
11	-38	-37	-36	-45
12	-38	-39	-40	-46
13	-37	-43	-45	-47
14	-37	-45	-48	-48
15	-37	-45	-45	-49
16	-40	-46	-52	-52

Table 2b.- Minimum Temperatures in January and February.

Height km	Alaska °C	Northern States °C	Southern States °C	England °C
1	-34	-36	-25	-9
2	-33	-37	-23	-15
3	-33	-43	-27	-21
4	-40	-46	-32	-29
5	-48	-48	-38	-38
6	-51	-51	-42	-46
7	-57	-56	-47	-51
8	-61	-60	-52	-56
9	-65	-66	-59	-58
10	-71	-71	-65	-63
11	-74	-74	-67	-68
12	-69	-76	-69	-72
13	-69	-79	-73	-70
14	-66	-76	-80	-66
15	-65	-73	-82	-66
16	-67	-77	-87	-68

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1. London, Meteorological Office. Observatories Year Book, 1926 - 37.

2. London, Meteorological Office. Temperature of the upper air over England. M.O.M. 406. 1941.

√ 3. Washington, Weather Bureau. Upper air maximum temperatures at standard levels over United States and Alaska. Technical Note. Washington. 1943.

√ 4. Washington, Weather Bureau. Upper air minimum temperatures at standard levels over United States and Alaska. Technical Note. Washington. 1943.

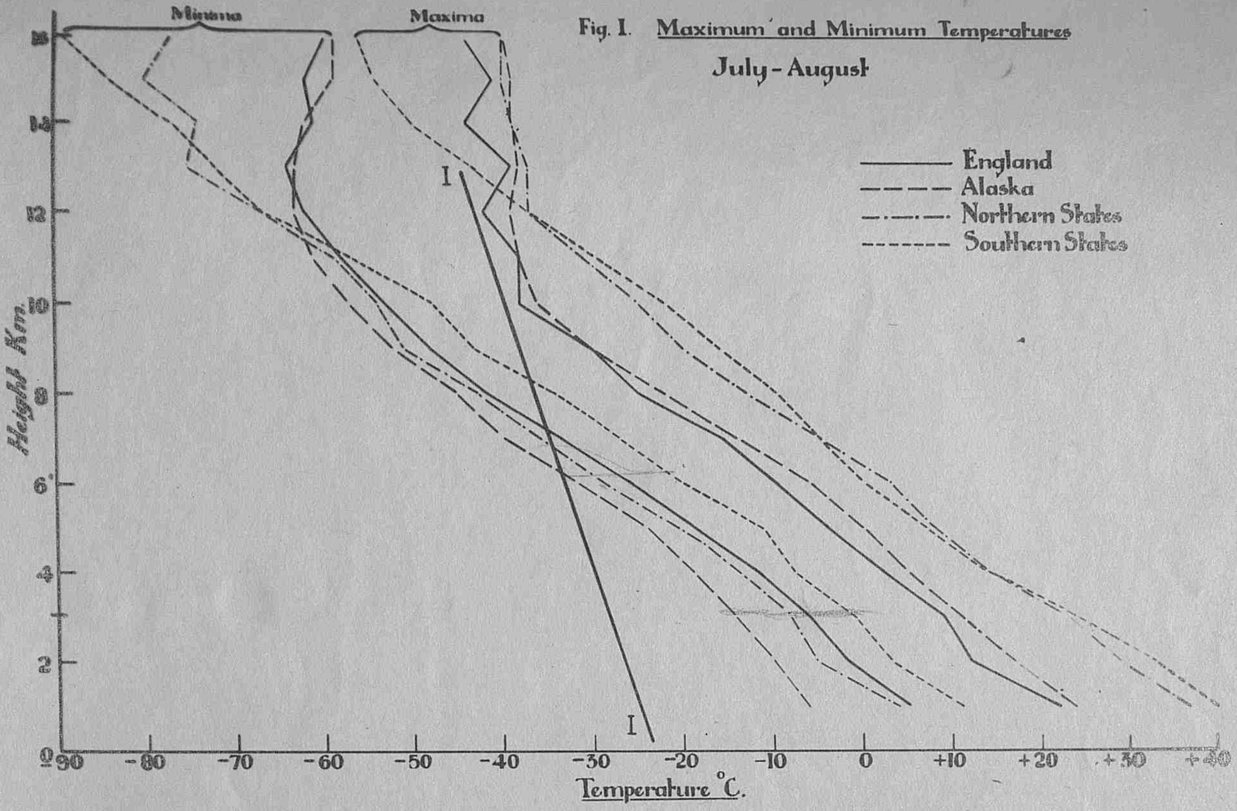

Fig. I. Maximum and Minimum Temperatures

July - August

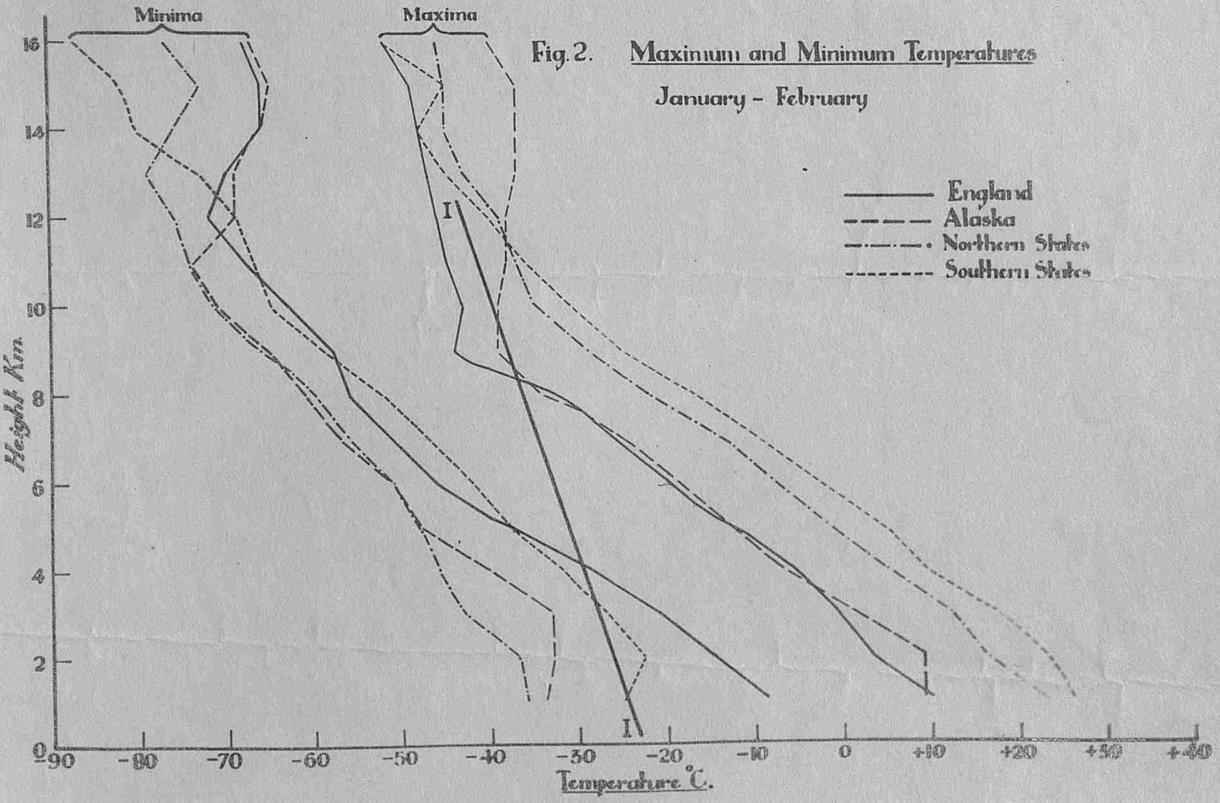

Fig. 2. Maximum and Minimum Temperatures

January - February

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by authority of M.P.P. 305 12 Oct 1945
by date 16 Apr 5

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AIR MINISTRY

M.R.P.155
25th November, 1943.

Gr. Brit — METEOROLOGICAL RESEARCH COMMITTEE.

REPORT ON EXISTING INFORMATION RELATING TO GUSTINESS.

m 51

By A.H.R. Goldie.

INTRODUCTION: The question of gustiness in the atmosphere, as bearing on the design of aircraft, was discussed at a meeting of the Structure Sub-Committee of the Aeronautical Research Committee, held on 22nd October 1943. After considering existing information on the matter, the Sub-Committee asked that the problem of diect measurement of gust structure in the upper air should be referred to the Meteorological Research Committe, adding that a recording instrument capable of being used on transport aircraft might be advantageous. The question was raised at the meeting of the Meteorological Research Committee held on 6th November 1943, when the Committee asked for a report to be made on the existing information relating to gustiness.

This report deals first (A) with wind fluctuations near the ground, then (B) with variations, particularly in the form of vertical currents, at higher levels, and finally (C) with the further information which seems to be required.

A. Wind Fluctuation near the Ground.

It is convenient to recognize three principal types of wind fluctuation - (1) Disturbances of short period, up to perhaps 10 seconds, set up in the main air stream by friction with the ground and obstacles. Except perhaps when they are very big these are of the nature of eddies and have their axes in all directions thus giving an isotropic distribution of wind variations. These eddies cause the speed and direction records of anemographs with the normal time scales to appear as a more or less uniform band or ribbon rather than a line. (2) There are larger disturbances rather in the nature of convectional cells or upsets of masses of air, drifting along in the main current. They show up as irregularities in the anemograph ribbon. (3) In addition to the above disturbances, there are more-or-less sudden shifts of wind associated with discontinuities.

The Cardington Report[2] (dealing with wind structure over level country in not very strong winds) gives the dimensions of eddies of

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the first type as of the order of 50 to 100 feet* in diameter and regards them as quite distinct from disturbances of the second type where the cell dimensions, horizontally, are of the order of 3000 to 8000 feet.

In the case of gales, however, it would appear that the eddies of the first type must be much larger. In the Bell Rock report[3] it is shown that the eddies have probably the dimensions of sea waves - up to perhaps 500 or 600 feet and that the motion is greater in the direction of the wind than transverse to it.

Probably the greatest dimensions of gustiness of short period recorded by anemometers in this country are those shown in the Paisley Observatory anemogram for the great gale of January 1927. This gave an extreme range in wind speed of about 140 f/s but it is not possible to say how many seconds elapsed from lull to maximum velocity, - probably at least 10 or 20 seconds and not more than two or three minutes. (On this occasion two tram-cars, each weighing 20 tons, were blown over on their sides.)

This kind of gustiness which is of the type generated over the housetops of a town in a gale of exceptional severity is of course most exceptional.

Among ranges of gustiness in gales at sea may be quoted 85 f/s in a wind with gust extremes up to 100 m.p.h. at Bell Rock.

There is not much information about fine structure of gust profile, since the inertia of instruments designed for continuous use and normal meteorological purposes renders them unsuitable for recording wind changes taking place in one or two seconds or even fractions of a second.

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* According to some estimates which I made from the Cardington "ultra-quick" run records, the sizes of the eddies were of the following order:-

Speed (M.P.H.)	13-20	26-40
Eddy (ft.)	136	176

A few special records, however, have been made. Simmons[4] quotes a case from hot wire anemometer records of a fall from 75 to 45 f/s in 0.1 seconds and a rise again to 68 f/s in 0.32 seconds.

Sherlock and Stout[5], Michigan, with a special moving plate anemometer recorded a rise of 17 f/s in ½ second in a wind of 40 to 50 m.p.h., i.e. in a distance of some 35 feet travel.

Examination of the Cardington "ultra-quick" runs shows a fall of 27 f/s in 2 seconds, equivalent to 13.5 f/s^2 in the speed record, but the direction record shows changes at the rate of one per second, so that if the fall just quoted actually occurred within a second the rate becomes 27 f/s^2.

As to eddies produced by local topography, Koschmieder's measurements[6] at the Wasserkuppe by no-lift balloons gave 88 and 140 metres for diameters of eddies which "probably extended further", the centres travelling 5 and 7 m/s respectively. Above a critical velocity of about 10 m/s these eddies disappear and the air current takes on its general direction with generalized turbulence. The "Relation between Ground Contour, Atmospheric Turbulence, Wind Speed and Direction" is discussed very fully by Morgans[7].

B. GUSTINESS at higher levels.

The type of gustiness chiefly dealt with in Section A above falls off considerably with height.

J.S.Dines[8] (from observations of the pull of a kite) estimated that at 1000 metres and over, the horizontal gustiness factor is only some 30% of that at ground level.

Proximity to the ground must limit vertical motion, but at appreciable altitudes above the ground vertical currents due to convection are probably more important than the fluctuations in horizontal air speed. They are presumably also of greater importance to the designer of aircraft, so this part of the question will be discussed in greater detail.

The existing information about vertical currents comes chiefly in three ways: -

(a) From measurements of actual rate of rise of pilot or registering balloons as compared with the rate by theoretical formula or from rate of rise or fall of "no-lift" balloons.

(b) From measurements with gliders.

(c) From vertical accelerometer records taken in aircraft or registering balloons.

Pilot balloons

Using the first method J.S.Dines[9] in 1913 found that upward currents tend to be smaller in cross section but greater in intensity than down currents. Convectional currents frequently attain their maximum intensity about 1000 feet above ground and fall off rapidly above 2000 feet; the largest measured by him was 16 f/s under st.cu cloud.

Dobson[10], in observations made in 1914 at Upavon with "no-life" balloons found upward currents of about the same order of magnitude and also that eddy motion, at heights of about 200 to 1000 feet, was roughly equal on the average, along and transverse to the wind and vertically.

In India in 1935 Ramanathan and Ramakrishnan,[11] using "no-lift" balloons released at heights from 0.5 to 2.0 km., found much the same downward and upward values as in England - some 15 f/s near cu clouds.

Chadburn[12] at Suffield in a longer series of observations from July to November 1942 found extremes of the order of 30 f/s upward and 15 f/s downward "in the absence of phenomena such as thunderstorms".

Since the above experiments depended on seeing the balloon in clear air they give no information about what happens inside clouds.

Glider Results.

Three years ago Miss Sawyer[13] extracted the available information obtained in gliders in and around clouds. For the most part the upward speeds recorded do not exceed 30 f/s and the downward speeds 20 f/s. There are, however, three exceptions:-

(I) Dittmar[14] in his world record flight in South America estimated that at times the lift was from 35-75 f/s.

(II) Kronfield[15] at a height of 20,000 feet in Cumulus estimated an upward current of 75 f/s.

(III) Koschmieder[16] states that in the July 1938 Rhon competitions glider ascents at mean rates of 65 f/s were made through thunder clouds several km. deep. From this he considers that in small regions in especially violent thunderstorms 100 f/s or more may occur.

At this point reference may be made to deductions made from falling waterdrops. Lenard's[17] results showed that no rain can fall through an ascending current of air whose vertical velocity is greater than 26 f/s. There is therefore good reason to believe that currents considerably exceeding this velocity occur frequently in hailstorms.

Accelerometer Records from Aircraft

In an investigation of Atmospheric Turbulence by Aircraft Carrying Accelerometer - Jennings, Alston and Howarth[18] in 1932 reached the following conclusions - "In S.E. England the maximum and minimum values of the normal accelerometer readings recorded were 2.3 g and 0.3 g respectively, and both occurred when flying in the neighborhood of cumulus clouds. On the hypothesis of a simple sharp edged gust the corresponding gust velocities are 17 ft/sec upward and 9 ft/sec downward. Owing to the absence of suitable weather conditions it was not possible to carry out a useful systematic investigation of the effect of speed and wing loading on the normal acceleration.

There is little likelihood of meeting bumps of greater severity over S.E.England, except perhaps in types of weather not experienced this summer, e.g., in abnormally hot or abnormally squally weather.

Further tests carried out in the Northern Highlands produced maximum and minimum accelerometer readings of 3.0 g and 0 g respectively. The corresponding vertical gust velocities are 30 ft/sec upward and 15 ft/sec downward respectively." ("If the more reasonable assumption is made that the entry to the gust is not abrupt but that the vertical velocity increases from zero to a maximum over some finite distance, then the maximum velocity required to produce the same distrurbance will be greater than the above.")

More recently a number of papers have been published by the N.A.C.A.[19] describing recent work in the U.S.A. These papers have been discussed and summarized by S.V.Fagg[20]. Quoting from his summary - "The first problem is naturally that of determining the magnitude of atmospheric gusts. The method so far used has been to measure normal accelerations due to gusts, together with the associated airspeeds, and thence to calculate the gust speed U_e from the sharp-edged formula:

$$U_e = \frac{\Delta n\ W}{\frac{1}{2}\rho_o V_1 S \frac{dC_L}{da}}$$

U_e = effective gust speed (E.A.S.)*
Δn = acceleration increment due to gust
w = weight of aircraft
ρ_o = sea level air density
V_1 = aircraft speed (E.A.S.)
S = gross wing area
$\frac{dC_L}{da}$ = slope of wing lift and incidence curve

* The expression true effective gust speed (U_{et}) is also sometimes used. U_{et} bears the same relation to U_e as true airspeed bears to indicated airspeed. True gust speed (U_t), as mentioned later, needs to be further computed from the aircraft size and wing loading and an assumed or estimated gust gradient distance (H), also referred to later.

Most of the work in measuring U_e has been done with V.G. recorders. These are installed in the aircraft, and describe a continuous record of normal acceleration against indicated airspeed. Over a period of time an envelope of maximum accelerations and speeds is thus obtained. Apart from the necessity of making some assumption with regard to the weight of the aircraft (e.g. take-off weight or mean weight) the record gives the information necessary for calculating the maximum effective gust speeds encountered. The simplicity of the V.G. recorder lends itself to wide application. In 1937 Rhode[19] recorded the V.G. results of 20,000 hours flying on over-land and ocean air transport, and a further note by Walker[19] in 1942 extends the results to cover 134,000 hours of V. G. records.

The next step is to deduce true gust speeds from the effective gust speeds observed. The ratio of true gust to effective gust as given by Rhode depends upon the aircraft size and wing loading, and on the gust gradient distance H. (This is the horizontal distance in which the linear gust reaches its maximum, after which the gust is assumed constant). When V.G. recorders are used it is necessary to assume a value of H. An improved method, though not capable of such wide application, is to take separate readings of airspeed and acceleration on a time base. The gradient distance can then be deduced from the time taken for the acceleration to reach its peak, assuming that this occurs at the same time as the gust reaches its maximum. This method is being used in the gust research being carried out on the XC.35 aircraft[19]. With this aircraft traverses are made through suitable cloud formation, when these are available, to explore their turbulent structure. Bulletins on the results are issued periodically.........

The Relation between true gust and effective gust speed..............

This is important since the normal acceleration at the C.G. (from which the effective gust is calculated) is the quantity most easily and most frequently measured, while the true gust speed is the fundamental characteristic of the atmosphere..... Assuming a value of H, Rhode deduced true gust speeds up to 80 f/s from the early V.G. records, which were obtained on aircraft of low wing loading......... Bulletins issued so far on the XC35 gust research have shown effective gusts up to 35 f/s E.A.S. and true gusts up to 85 f/s true speed......

The gust gradient distance is usually obtained from the time to reach maximum acceleration, assuming that the acceleration reaches its peak at the same time as the true speed. The XB. 15 bomber tests indicated that the gusts giving the greatest acceleration are those with gradient distance H of the same order as the span........

One point which has emerged from the gust work is the desirability of investigating the lateral extent of gusts."

Fagg comments that the method of deducing gust speed by measuring acceleration on aircraft is not entirely satisfactory owing to the large number of factors involved.

A further N.A.C.A. report[20] on a "Preliminary Analysis of the lateral distribution of gust velocity along the span of an airplane" states that:-

"About 80% of all the gusts measured were of a roughly triangular shape, i.e. they reached a peak at a point along the aircraft span and fell off in intensity to each side. The 'rectangular' gust - equal intensity across the span - as at present used for design, occurs relatively infrequently. Lateral gust gradient distances were calculated*; and the values ranged from 20 feet to 300 feet. The average (or most probable) lateral gust distance was about 80 feet (about 9 chord lengths) for all values of the gust velocity.

As a result of this analysis the suggestion is made in the U.S.A. report that it would be better to replace the present symmetrical design gust (30 K ft/sec rectangular)# by a gust which has a value 36 K ft/sec. at the aircraft centre line and falls off to 24 K ft/sec at each wing tip."

Donely[21] using a series of measurements made in a light airplane, discussed the maximum gust intensity in relation to gradient distance (H) and to various meteorological quantities. He found little or no correlation between U and H though dot diagrams which he reproduces suggest that on the whole the gust velocity increases with the gradient distance and that the Aeronca machine used responds most frequently to gusts having gradient distances of the order of 30 feet.

He found little or no correlation between either U or H and various meteorological quantities (potential temperature θ or $\frac{d\theta}{dh}$, wind w or $\frac{dW}{dh}$).

#K is defined in XC.35 Gust Research Project. Bulletin No.6 (A.R.C. 6018) as "the effective gust factor for the important gust gradient distance of nine chord lengths (from unpublished data)" and a reference is given to U.S. Civil Air Regulations 04 - Airplane Airworthiness - as amended to May 31, 1938. K depends on wing loading. 30 K is equivalent to 35 f/s at wing loading of 40 lbs.sq. ft. The British design gust, also sharp-edged, is 25 f/s. The overall safety factor in both countries is 1.5 up to maximum diving speed.

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* Presumably from the readings of airspeed and acceleration against time, the gradient distance being deduced from the time taken for the acceleration to reach its peak, assuming that this occurs at the same time as the gust reaches its maximum.

He also tried various relationships involving Richardson's number $\frac{g}{\theta}\frac{d\theta}{dh} / \frac{(dW)^2}{(dh)}$. Correlation obtained was negligible but the result might have been due to limitations of the data.

Accelerometer Records from Registering Balloons

A few observations on turbulence in the high atmosphere were made by Junge[22] from measurements of non-uniformity of rate of ascent of registering balloons. An instrument was used which recorded acceleration and pressure, together with the temperature of the instrument itself. The last mentioned was required for correcting the recordings to a standard basis. The author remarks:-

"We can say the following about the recordings in general: We see from Fig. 5 that accelerations of more than 0.3g are hardly ever exceeded; but according to Table P.106 that means oscillations of more than 20° rarely occur, and therefore with homogeneous turbulence the horizontal accelerations of the balloon would hardly invalidate the recordings. - In comparing the turbulence at the different heights we must take into consideration the following: As the recording of the inside temperature shows, the temperature of the apparatus in the stratosphere has become about -40° in spite of the black coverings. According to Fig.6 that corresponds to a decrease in the sensitiveness of the acceleration-indicator of about 20% as compared with that at room-temperature. Furthermore the heights marked on the recordings show that as a result of the loss of gas the rate of ascent decreases upwards. This again - according to P.106 causes a loss of sensitivity which however is hard to estimate. Therefore in a comparison between troposphere and stratosphere the deflections should be increased by about 20 to 30% in the latter. - The variation of the interior temperature of the apparatus reflects the outer temperature-variation so faithfully that we could deduce from this the approximate height of the tropopause.

Unfortunately there is insufficient material to enable us to draw general conclusions. But it is clear from the measurements that the accelerations in the stratosphere are greater than in the troposphere, especially the high troposphere. If we assume that the accelerations measured reflect on the whole the vertical accelerations of the air, which is very probable from the above remarks, then turbulence would be strongest in the stratosphere. This result is very well confirmed by the double soundings carried out in Lindenburg in 1936 which came into my hands shortly before this work was finished. Three of these went high up into the stratosphere. In Fig.14 the values of the rate of ascent are given (at minute intervals). The great increase of the variations in the stratosphere is so striking that there is no question of explaining it by inexact observation as the balloon goes farther away. This result, together with the acceleration-recordings, shows that the small-scale turbulence of the air, which actually causes the acceleration of the balloon, as well as the large-scale turbulence, is just as great if not greater in the stratosphere than in the troposphere."

C. Further information required.

It appears that from the point of view of the aircraft designer the chief gaps in the meteorological information on gustiness relate to:-

(a) the maximum vertical speeds occurring in the atmosphere, particularly at levels above the first kilometre;
(b) if speeds occur above those at present allowed for in design, the relative frequency and the conditions attending such occurrences;
(c) the lateral extent of vertical gusts or the variation of the vertical speed in the horizontal direction.

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AIR MINISTRY

MRP 156 (Revised)
7th April, 1944.

(See MRP 162)

METEOROLOGICAL RESEARCH COMMITTEE

Note on the Errors of Cup Anemometers in Fluctuating Winds.

by F.J.Scrase and P.A.Sheppard.

[This note includes comments made on M.R.P.156 by P.A.Sheppard, M.R.P.162]

1. The purpose of this note is to place on record some information which has been obtained on the extent to which cup anemometers over-estimate the mean speed of a gusty wind and on the differences, in this respect, shown by conical and hemispherical cups. We are indebted to the Meteorological Section, Ministry of Supply, Porton for the help given in carrying out the wind tunnel tests described below.

2. Some theoretical data on the question of over-estimation were included in the published description of the Sheppard anemometer (J.Sci.Inst.17, 1940, 218). The data were computed from theoretical results developed by O. Schrenk (Zeit. Techn. Phys.10, 1937, 57). In the equation of motion of a cup system

$$\frac{I}{R}\frac{du}{dt} = M - F$$

Schrenk assumes the following approximate expression for the aerodynamic couple M:-

$$\frac{M}{\frac{1}{2}\rho u^2 r^2 \pi R^2} \doteqdot \lambda\left(\frac{v}{u}\right)^2 + \mu\left(\frac{v}{u}\right) + \nu$$

where I is the moment of inertia of the cup system, R the cup arm length, u the circumferential cup speed, F the frictional moment, ρ air density, r the cup radius (for hemispherical cups), v the mean wind speed and λ, μ and ν are constants determined from measurements of the torque required to turn the anemometer at various speeds in a steady air stream. By integration of the equation and consideration of extreme cases, Schrenk derived a family of curves which, to a first approximation, shows the dependence of the inertia errors on the gustiness (expressed by the ratio of the amplitude of the velocity fluctuations to the mean wind speed $\Delta v/v$,) and on the dimensionless quantity K given by

$$K = 0.55\,\rho R^2 r^2 T v / I,$$

where T is the period of the velocity fluctuations (the technical system of units, m.,kg.,sec, are used in this expression). Schrenk's curves are reproduced in Fig. 1, the left-hand vertical scale giving errors expressed as a percentage of the mean velocity of sinusoidal fluctuations and that on the right giving the errors expected for the irregular fluctuations that occur in practice. The extreme cases occur (a) when K is zero i.e. either when the anemometer has infinite inertia and (b) when K is infinite, i.e. when the anemometer has no inertia.

3. Owing to a numerical error the computed data given for the percentage over-estimation of the Sheppard anemometer in the paper referred to above are incorrect. The correct values are given in the table below together with corresponding values for the standard Meteorological Office pattern cup anemometer, the dimensions of the anemometers being:-

...

	Sheppard Anemometer	M.O.pattern anemometer.	
R	0.07	0.16	metres
r	0.025	0.0635	metres
I	1.1×10^{-5}	165×10^{-5}	$kg.m.sec^2$
ρ	0.125 kg.sec^2/m^4		

Over-estimation of Sheppard and M.O.Anemometers computed from Schrenk's analysis.

v (m/sec)	2			5			10		
T (sec)	1	10	60	1	10	60	1	10	60
$\Delta v/v$	0.2	0.4	0.6	0.2	0.4	0.6	0.2	0.4	0.6
K for Sheppard	.038	.38	2.28	.09	.95	5.7	.19	1.9	11.4
K for M.O.pattern	.009	.09	.52	.02	.21	1.29	.04	.43	2.58
Over-estimation (% of v) for Sheppard	1.4	3.9	1.8	1.4	2.3	0.6	1.3	1.2	0.1
for M.O.pattern	1.4	5.7	7.8	1.4	5.0	4.0	1.4	3.8	1.5

It will be seen that these figures indicate practically no difference between the errors of the two anemometers for small fluctuations of short period and that the Sheppard instrument is less subject to over-estimation than the M.O.pattern for large and long period fluctuations.

4. In order to obtain some experimental evidence of the extent of the over-estimation some comparisons of Sheppard and M.O. pattern anemometers in steady and fluctuating winds were arranged to be carried out in the wind tunnel of the Meteorological Section, Porton. According to these measurements the Sheppard anemometer showed considerably less over-estimation than Schrenk's theory indicated. Thus for a wind speed of 4.75 m /sec., T = 1.5 sec. and $\Delta v/v = 0.36$, the observed over-estimations averaged 0.5% (actual errors ranged from -1.5 to +2.5%), whereas the computed value was 4.5%. The standard M.O.pattern anemometer, on the other hand, gave results more in accordance with Schrenk's theory; for a wind speed of 4.0 m./sec., T = 1.0 sec. $\Delta v/v = 0.25$, the observed over-estimation was about 3.5% as compared with 2.5% computed.

5. It was suggested by the Meteorological Section, Porton that the closer agreement of the M.O.anemometer with Schrenk's theory was accounted for by the fact that Schrenk used a hemispherical cup form similar to that of the M.O.anemometer whereas the Sheppard instrument has a conical cup form. Some further experiments were therefore made with an M.O. anemometer fitted with conical cups of exactly the same form as those of the Sheppard instrument but having the larger dimensions of the standard cups. It was then found that there was practically no difference between the M.O. anemometer and the Sheppard instrument under approximately the same conditions as before, i.e. short period fluctuations of wind.

6. On account of the gross empiricism of the aerodynamics of the cup anemometer one can only expect Schenk's analysis to be formally correct; the numerical constants in his work are likely to vary as soon as any parameter whatever of the cup system is changed, e.g. cup form, ratio r/R, beaded or unbeaded edge. As far as the design of the cup anemometer is concerned, Schenk's quantity K may be written,

$$K \propto R^2 r^2 / I$$

(If the cup is not hemispherical in form r may be defined as an /

equivalent radius of the cup and will be of the same order of size as the radius of the cup face).

Thus, $K \propto \dfrac{R^2 r^2}{M(G^2+R^2)}$ where G = radius of gyration of cup about a diametral axis,

$\propto \dfrac{R^2 r^2}{2\pi r^2 \sigma t(\frac{2}{3}r^2+R^2)}$ σ = density of cup material,

t = thickness of cup material,

$\propto \dfrac{1}{\sigma t(\frac{2}{3}\frac{r^2}{R^2}+1)}$

Since the ratio r/R is appreciably less than unity, K, and therefore the extent of over-estimation, is practically independent of this geometrical factor and of r and R separately. In a given cup it depends mainly on the product σt, i.e. the mass per unit area of the cup. For the Sheppard and the M.O.pattern anemometers the mass per unit area of the cups is approximately 0.15 and 0.59 gm./cm^2 respectively. Reference to Fig.1 and Schrenk's expression for K (page 1) suggests that this difference might cause appreciable difference in the amount of over-estimation of wind speed for the intermediate range of K (intermediate values of the product Tv) and relatively large fluctuation ratios.

7. Our general conclusions are, firstly, that although Schrenk's theory gives reasonably good values for the over-estimation errors of hemispherical cup anemometers it gives excessive values for conical cup instruments (this, however, is no criticism of Schrenk's treatment), secondly, that conical cup anemometers over-estimate the mean speed of a gusty wind to a much less extent than do hemispherical cup instruments. It was not practicable to extend the experimental comparisons to high wind speeds but since the driving force increases with wind velocity more rapidly than the inertia forces, the over-estimation should become relatively less important at high speeds - as is indeed indicated by Schrenk's expression for K and fig. 1.

8. Some earlier comparisons of hemispherical and conical cups in steady winds had shown no particular advantage in the conical form apart from a tendency for the conical cups to start moving at slightly lower speeds. The difference was so slight, however, that at the time it was not thought worth while changing from the hemispherical form in the case of the M.O.pattern anemometer. In view of the results discussed above, however, it has been decided to adopt conical cups, similar in form to those of the Sheppard instrument, as standard for M.O. cup anemometers.

MRP 156

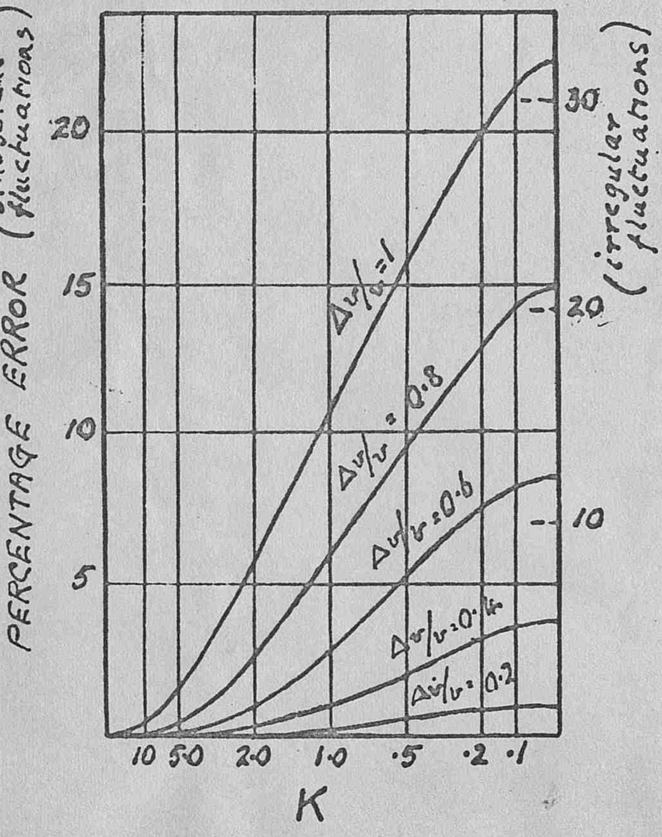
PERCENTAGE ERROR (Sinusoidal fluctuations)
5
10
15
20
(irregular fluctuations)
10
20
30
Δv/v = 1
Δv/v = 0·8
Δv/v = 0·6
Δv/v = 0·4
Δv/v = 0·2
10
5·0
2·0
1·0
·5
·2
·1
K

FIG. 1

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[illegible] #141

MRP 136 Revised

SOME RECENT DEVELOPMENTS IN METEOROLOGICAL INSTRUMENTS.

By F. J. SCRASE, M.A., Sc.D., F.Inst.P., Superintendent of Instruments, Meteorological Office, Air Ministry

Introduction

The requirements of meteorological instruments differ to some extent from those of the instruments of the physical laboratory. In the first place most meteorological instruments have to be capable of withstanding long exposure to weather conditions of all kinds; in the second place many of them have to be in continuous operation day after day for years on end with little more attention than is required for the changing of recording charts. Further, many meteorological instruments have to be handled by amateur observers who, although competent to make very useful contributions to the collection of meteorological data, may not possess the physical knowledge or mechanical skill necessary to operate complicated instruments such as are found nowadays in the physical laboratory. Meteorological instruments must, therefore, generally be robust and simple in design, and since many of them fulfil these requirements reasonably well their designs have remained practically unchanged for many years.

Most of the developments in meteorological instruments that have taken place in recent years are not so much radical changes in design as the adaptation of existing designs to give remote indication or remote recording. In general it is desirable for meteorological measurements to be made in a well-exposed position away from the effect of buildings, but it is very convenient if the measuring instruments can be arranged to indicate or record their readings in the meteorologists' central office. An especially important development of the distant indication idea has resulted from the application of radio-telegraphy to meteorological instruments for sounding the upper atmosphere, and in view of the importance of this work a considerable part of this article will be devoted to instruments of this kind.

Instruments for measuring wind velocity and direction

The instrument that is most commonly used for measuring wind velocity is the cup anemometer. It consists of three or four cups attached to the ends of arms which are pivoted so that they can rotate in a horizontal plane. The difference of wind pressure on the convex and concave surfaces of the cups causes the arms to spin round. One of the advantages of this type of instrument is that it is independent of the direction of the wind and it does not need to be orientated or mounted on a direction vane.

It is usually assumed that the speed of the cups is strictly proportional to the speed of the wind but this was not true for the earlier patterns of cup anemometers, there being a tendency to read too low at low speeds and too high at high speeds. In recent years, however, much attention has been paid to the design of these instruments with a view to obtaining better performance. The first improvement was the substitution of the earlier arrangement of four cups on arms spaced 90° apart by three cups on arms spaced at 120°. This improvement was developed by Patterson (1926), Director of the Canadian Meteorological Office. It was followed by the introduction by Marvin (1934), of the United States Weather Bureau, of cups with circular beading round the edges. With these two modifications the factor of the cup anemometer is practically constant over a range of wind speed from 5 m.p.h. to about 100 m.p.h.

Observations of wind speed with cup anemometers are usually made by means of a dial counter built into the base of the instrument or by means of electric buzzer signals which are made by a contact maker geared to the spindle of the cup system. An instrument of the latter kind is illustrated in Fig. 1. Cup anemometers of these two patterns give the run of the wind in a measured interval of time and they are unsuitable for the measurement of instantaneous wind speeds. For this purpose instruments in which the cups are arranged to turn the permanent magnet of a small electric generator are available. The indicator in this case is a voltmeter calibrated in terms of wind velocity. This type of instrument is well adapted for direct indication of wind speed at a moderate distance from the place of measurement, but it is not sensitive at low speeds.

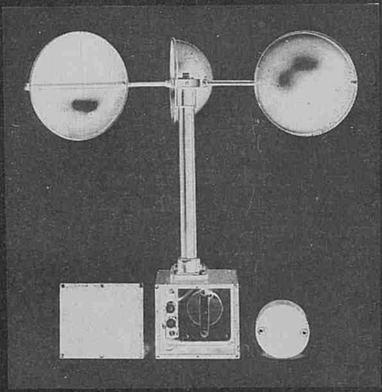


Fig. 1. Electric cup anemometer

A cup anemometer that is especially suitable for the measurement of low wind speeds has been developed recently by Sheppard (1940). It is illustrated in Fig. 2 and is appreciably smaller and more portable than the usual type of cup anemometer. Its chief feature is the employment of a counting mechanism consisting of a continuous motion stop-watch from which the hair spring and balance are removed. The watch lever is linked to an eccentric on the anemometer spindle which merely performs a trigger action to which the watch mechanism responds with the energy supplied by its mainspring. A three-cup system is used for this instrument but the cups, which have beaded edges, are conical instead of hemispherical in shape. It is claimed that the instrument gives a linear relationship between the cup speed and the wind velocity down to velocities as low as 0·5 m.p.h. Another advantage which this instrument has over the more usual design of cup anemometer is that owing to the small inertia of the moving parts the system is very sensitive to changes in wind velocity and the tendency to overestimation in a gusty wind is negligibly small. Unlike the larger instruments, however, the Sheppard instrument in its present form is not intended for continuous use over long periods; it may be regarded as replacing the vane airmeter to a very large extent.

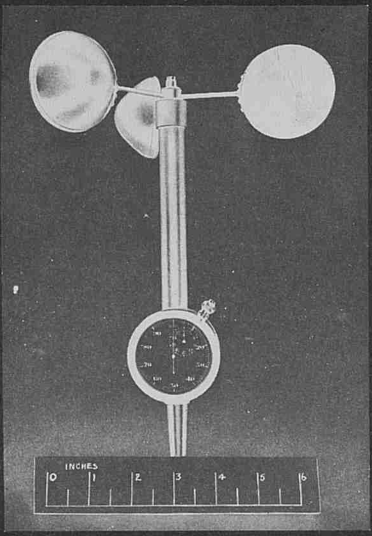


Fig. 2. The Sheppard anemometer

For obtaining continuous records showing not only the mean direction and velocity of the wind but also the rapid fluctuations the pressure tube anemograph is still one of the most reliable instruments. In this instrument the direction vane incorporates pressure and suction apertures which are connected by pipes to a cylindrical vessel of water containing a float, the height of which is proportional to the wind velocity and is recorded on a rotating chart. To a very limited extent the instrument is a distant recording one, since the velocity head can be about 100 ft. from the recording gear, but it is usual to have the recorder vertically below the head so as to obtain a direct coupling between the vane and the direction recorder. This latter is a cam mechanism that transforms the angular motion of the vane into linear motion on the chart.

Since it is sometimes desirable, especially on aerodromes, to have the recorder inside an office building and the velocity and direction head several hundred yards away where it can be well exposed and at the same time not in a dangerous position for aircraft, a means of making the apparatus truly distant recording was developed by the Meteorological Office (see Dines, 1937). Use is made of small synchronous electric machines which go by the name of Selsyn units in this country and Autosyn units in the United States. These units have a three-phase field winding and a single phase armature; when they are suitably connected in pairs with the armatures fed by the same a.c. supply as shown in Fig. 3 any angular motion given to one unit is faithfully reproduced by the other.

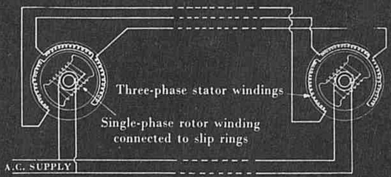

Three-phase stator windings
Single-phase rotor winding connected to slip rings
A.C. SUPPLY

Fig. 3. Connexions of a pair of self-synchronous motor units

The arrangement of a pressure tube anemograph for remote recording is shown in Fig. 4. For the recording of wind direction the application of the self-synchronous units is straightforward, the transmitting unit *T* being mounted below the direction vane with the armature coupled to the vane, while the distant receiving unit *A* is coupled to a direction recorder *B* similar to the standard pattern. It is essential that the receiving unit should give sufficient torque to move the recorder; medium sized Selsyn units are found to be adequate for this purpose.

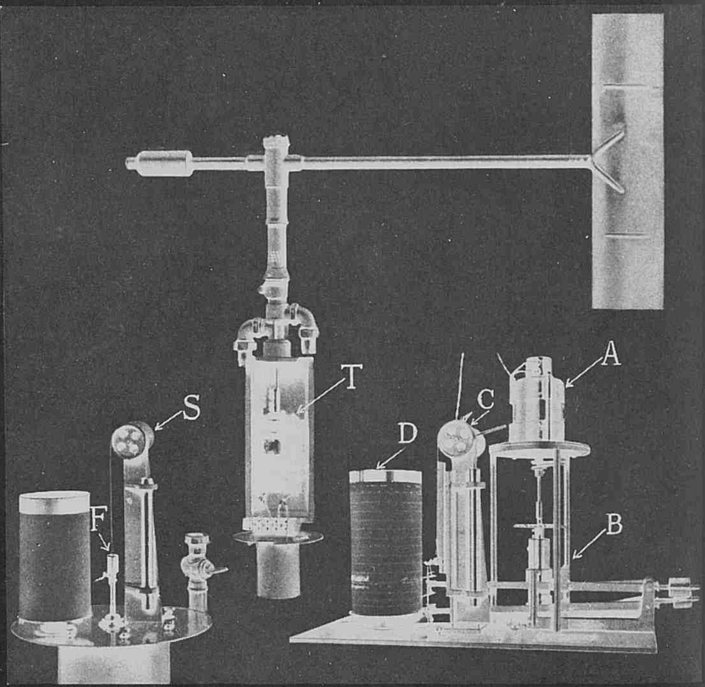


Fig. 4. Distant-recording pressure tube anemograph

In the case of the velocity record the linear motion of the float has to be transformed into angular motion of the transmitter armature. This is done by coupling the transmitter *S* to the float rod *F* by a light chain passing over a pulley attached to the transmitter which is mounted just above the float rod. Since the force moving the float in light winds is very small, it is essential that the transmitter should possess very little inertia; for this reason a smaller and lighter type of Selsyn unit is used. At the distant recorder a receiving unit *C* is arranged, in a similar manner, to re-transform the angular movement into the linear movement of a pen. A clock drum *D* carries the chart on which the record of velocity and direction is made. The errors of the distant recorder as compared with a direct recording anemometer do not exceed 1 m.p.h. in general.

An improved pilot balloon theodolite

The method of measuring the speed and direction of the wind in the upper air by observing the movements of small free balloons involves the use of a theodolite which differs from the normal surveying instrument in having a right angle telescope with the eyepiece tube forming the horizontal axis of rotation. With this arrangement the observer's eye is directed horizontally no matter how high the elevation of the object. Such theodolites have been in use for about 30 years, but during the last year or two a number of improvements have been made in the design of these instruments, mainly with a view to providing easier manipulation and reading.

The path of a pilot balloon is usually a very erratic one and in following a balloon with a theodolite the observer has to be able to take readings of azimuth and elevation very quickly, otherwise the balloon may drift out of the field of view and be difficult to sight again. In the earlier instruments the elevation circle was concentric with the horizontal axis of the telescope and not always sufficiently close to the azimuth circle to permit of rapid reading of the two scales. In the new pattern theodolite of the Meteorological Office (Cranna, 1938) illustrated in Fig. 5 the introduction of bevel gearing between the horizontal telescope barrel and the elevation circle enables the latter to run parallel to and immediately above the azimuth circle. With this arrangement both circles can be read through one window directly below the eyepiece.

The vertical and horizontal movements of the telescope are controlled through worm gearing by hand wheels. These are divided to indicate tenths of a degree and are placed one on each side of the window in a position that is convenient both for manipulation and for reading. A throw in and out mechanism operated by levers near the front of the instrument allows the telescope to be disengaged from the hand wheel gearing.

Another useful feature of this instrument is the provision of a subsidiary wide angle telescope having a field of view four times as large as that of the main telescope. The same eyepiece is used for both telescopes but a short focus objective is used for the wide angle; it is brought into operation by means of a hinged mirror which can be turned from one position to the other by a lever. The auxiliary telescope is very useful for sighting a balloon during the initial stages of an ascent, especially at night. The optical system of the main telescope differs from that of the earlier right angle telescopes in having a pentaprism instead of a triangular prism; a slight accidental displacement of a pentaprism has no effect on the horizontal collimation.

Built-in electric battery lighting is provided in this theodolite, a lamp being fitted to the front of the instrument for illumination of the scales and another lamp being attached to the telescope tube to illuminate the field of view by reflexion. The illumination of the field of view can be varied continuously from zero by adjustment of the reflecting surface.

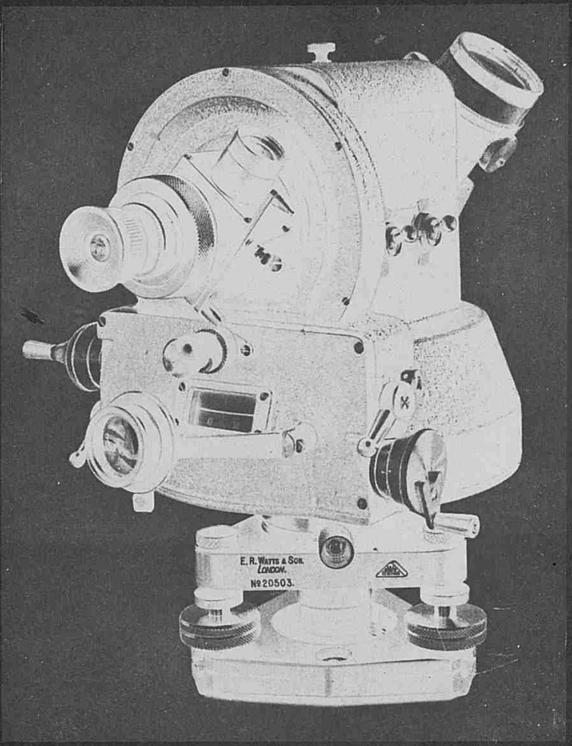


Fig. 5. The Meteorological Office pattern pilot balloon theodolite

An instrument for measuring visibility at night

The importance of visibility for the navigation of aircraft has led to a demand for more precise methods of measuring the degree of transparency of the atmosphere. In most cases, however, the instruments that have been devised for this purpose are too elaborate for general use where the time required for making an observation is an important factor. For this reason it is still the usual practice to estimate visibility in the daytime without the aid of an instrument. Objects are selected at a number of distant distances and the visibility, which is largely dependent on the degree of transparency of the air, is determined by noting the most distant of the series of objects that is visible. The method becomes difficult at night and to a large extent the observer is left to his own devices to obtain an estimate which is comparable with a daylight observation in air with the same degree of transparency.

In order to overcome this difficulty of estimating visibility at night, Gold (1939) devised a very simple form of visibility meter which, while not giving the precision of a more elaborate photometric method, nevertheless gives sufficient accuracy for general meteorological purposes. The instrument, which is illustrated in Fig. 6, makes use of the fact that the brightness of a constant source of light at a fixed distance depends on the transparency of the air and can be measured by the obscuring power of a screen or filter of sufficient density to reduce the light to the limit of visibility. The filter consists essentially of a neutral optical wedge *AB* about 10 in. long, whose density increases uniformly from one end to the other, against which is placed a similar but shorter wedge *C* having precisely the same gradation of density but in the opposite direction. The combined density of the two wedges is uniform over the area of the shorter one but it can be varied by sliding the short wedge in the frame in which the wedges are mounted.

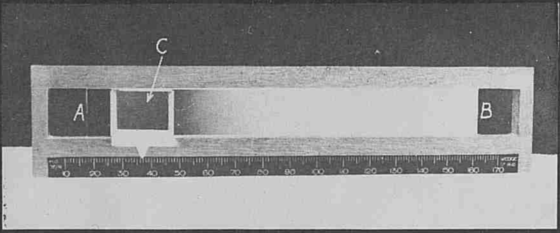


Fig. 6. The Gold visibility meter

The filter is graduated in units of obscuring power which have been given the name 'nebules'; the unit is defined such that 100 nebules reduce the intensity of light to 1/1000th part of its incident value. In using the instrument a source of light of known intensity at a suitable distance is selected and is viewed through the filter which is adjusted until the light is reduced to the limit of visibility. The obscuring power of the filter is then equal to the difference between the obscuring power necessary to render the light invisible in perfectly clear air and that of the air intervening between the light and the observer. The value for clear air can be determined for each light source in use and the scale readings of obscuring power can then be readily converted to values on the daylight visibility scale by means of suitable tables.

RADIO-METEOROGRAPHS

The meteorologist of the present day makes great use of any information he can obtain about conditions, especially temperature, in the upper air, and the collection of aerological data is now a very important part of a meteorological service. Before 1914 most of the information about the temperature and humidity of the upper air was obtained by sending up recording instruments attached to sounding balloons. The information recorded by these meteorographs was not available until the instruments were recovered; it was seldom, if ever, available in time to be of any use to the weather forecaster. In later years the forecaster's needs were served by observations made in aircraft, but such observations cannot provide information for great heights such as can be reached by unmanned balloons and moreover they are costly to make. The application of radio technique has now provided a means of obtaining upper air data from great heights without appreciable delay and at not unreasonable cost. The instruments used in this method are known as 'radio-meteorographs' or 'radio-sondes' and they have been developed along distinctive lines in a dozen or so countries.

The main idea of the radio-meteorograph is that a mechanism recording the meteorological conditions, e.g. pressure, temperature or humidity, should be arranged to vary some element of radio transmission which can be received at a ground station and interpreted in terms of the variation in the meteorological conditions. The radio-meteorograph is, in fact, half meteorological instrument and half miniature radio transmitter, and the differences between the various types that have been developed lie mainly in the methods by which the meteorological elements control the radio transmission. The devices employed for measuring the meteorological conditions are generally similar to those used in ordinary meteorological instruments. Thus an aneroid diaphragm is generally used for indicating the pressure, a bimetallic strip for the temperature and human hair for the relative humidity. The radio-meteorograph, complete with batteries for supplying the power to the transmitter, is carried aloft by a balloon of about 5 ft. in diameter. The balloon usually reaches a height of from 15 to 20 km. before bursting and in order to break the fall of the apparatus after the balloon has burst a parachute is attached. Reception of the observations transmitted by the instrument is usually made on a short-wave receiver arranged for automatic or semi-automatic recording.

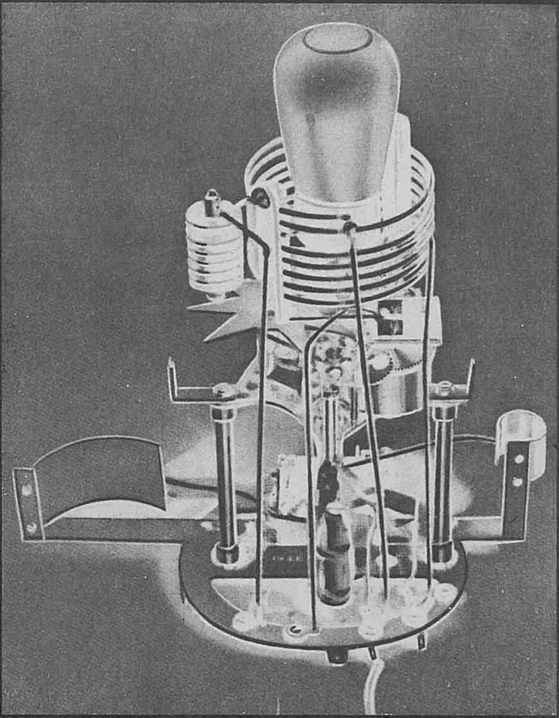


Fig. 7. The Bureau radio-meteorograph
(*Reproduced from 'Über Radiosonde Konstruktionen'*)

There are at least three ways in which the meteorological elements are made to control the radio transmission, viz. (*a*) by interruption of the radio signal; (*b*) by variation of the signal frequency; (*c*) by variation of the modulation frequency. Among the earliest instruments employing the first method are those developed by Moltchanoff (1936) and by Bureau (1935). In Moltchanoff's instrument the interruptions of the signal are made by an arm, rotating at a constant speed, which makes contact successively with movable pointers attached to the meteorological elements and with fixed reference pointers. Each contact completes the circuit of a valve transmitter which sends out a short signal. The time intervals between successive signals are a measure of the angular deflexion of the movable pointers which by calibration may be interpreted as values of the pressure, temperature and humidity.

In Bureau's instrument, which is illustrated in Fig. 7, the readings are made independent of the speed of rotation of the contact arm by arranging for the latter to modulate the frequency of the signal. This is done by including in the circuit a small plate condenser, the capacity of which is varied periodically by a toothed wheel geared to the contact arm. The angular deflexion of a pointer is then obtained by counting the number of cycles of modulation occurring between successive contacts. This is facilitated by passing the received signals into the coil of an oscillograph which is arranged to record in ink on a moving paper strip by means of a siphon-fed pen.

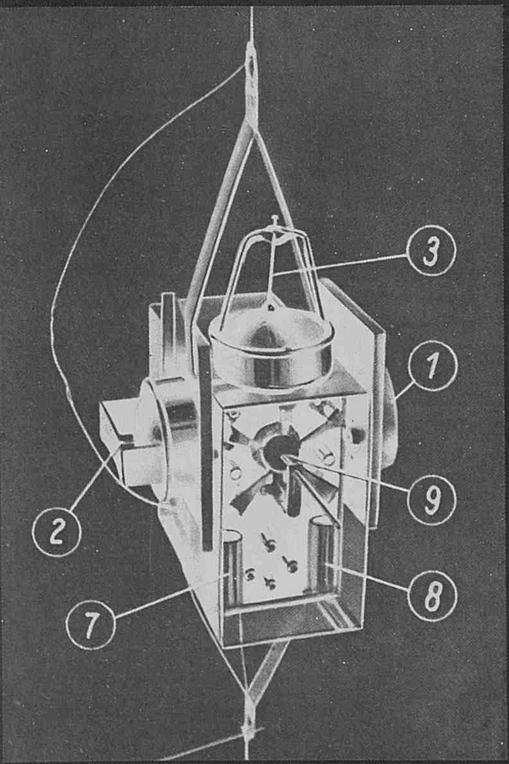


Fig. 8. The Väisälä radio-meteorograph
(*Reproduced from 'Über Radiosonde Konstruktionen'*)

Of the instruments in which the meteorological device is made to vary the radio-frequency, that designed by Väisälä (1937) is one of the most successful. This apparatus is illustrated in Fig. 8 and a diagram of the circuit is given in Fig. 9. In this instrument the thermometer element 2 (a bimetallic strip), the aneroid box 1 and the hair hygrometer 3 each control the capacity of plate condensers 4, 5 and 6. The three variable condensers together with two fixed condensers 7 and 8 are successively inserted in the oscillatory circuit of a transmitter by an air driven switch 9. Five signals of different frequency are thus emitted in turn and are measured on a semi-automatic receiver. The two fixed condensers are included in order to eliminate the individual characteristics of the transmitter and their capacities are selected so as to be at the ends of the range likely to be covered by the variable capacities. Since the frequency due to the fixed condensers does not vary much, the reception can be concentrated mainly on the signals which are varied by the meteorological elements. The switch that puts the condensers in circuit is operated by a very light windmill which rotates on account of the ascensional velocity of the balloon carrying the apparatus. Whereas most radio-meteorographs use dry batteries for the power supply to the transmitter, the Väisälä instrument uses light weight accumulators of the lead-acid type. Another noteworthy feature of the instrument is that it is extremely light, weighing only 365 g. including battery; most of the other types of radio-meteorograph exceed 500 g. and a few are very much heavier.

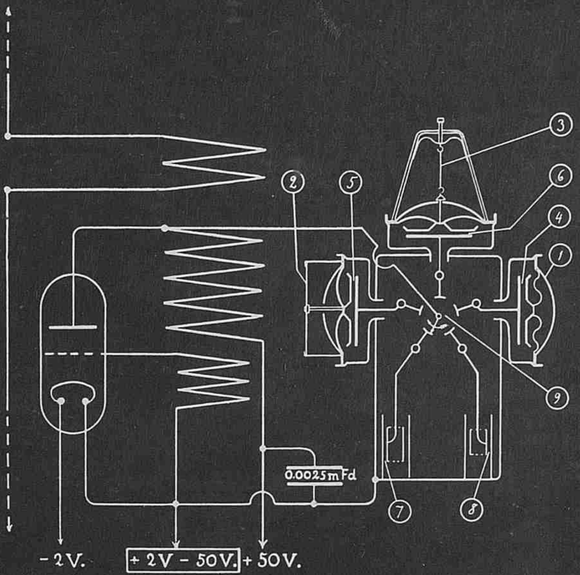


Fig. 9. Circuit diagram of Väisälä radio-meteorograph
(*Reproduced from 'Über Radiosonde Konstruktionen'*)

Another radio-sonde of the type in which the meteorological elements are arranged to vary the radio frequency is that used in the German Navy. The interesting feature of this instrument is that the use of mechanical switch contacts is avoided by having contacts built into the stem of a liquid-in-glass thermometer. The liquid used is a mercury-thallium alloy which remains liquid down to about −60° C. The contacts are arranged to vary the inductance of a transmitting circuit discontinuously at every 3° C change in temperature, but a condenser which is sensitive to temperature changes is included in the circuit so as to enable the continuous variation from one contact to the next to be followed. A somewhat similar arrangement is used for the pressure indications, a mercury-thallium index being arranged to close contacts in a gas-filled capillary tube and so causing a change, of short duration, in radio-frequency. The effect of temperature on the volume of the gas in the tube is allowed for when the pressure is derived from the measurements.

A radio-meteorograph in which the modulation frequencies are controlled by the meteorological elements was developed at the National Physical Laboratory for the Meteorological Office shortly before the war (see Thomas, 1938). A schematic diagram of the electrical circuit is shown in Fig. 10. In this instrument the barometer and thermometer units are arranged to vary the air gaps of iron-cored inductances which are included in low-frequency oscillatory circuits. The changes in inductance so produced vary the frequencies of the two low-frequency oscillators and the audio-frequency voltages generated by these oscillators are applied to a modulator which in turn controls the output of a radio-frequency oscillator. In the case of the barometer unit the air gap of the inductance is varied by the movement of a stalloy iron plate attached to an aneroid capsule. For the thermometer unit a fine brass wire stretched on an invar frame is used to control the movement of a similar plate. A photograph of the radio-meteorograph with the top cover removed is reproduced in Fig. 11; the shorter of the two cylinders contains the transmitter batteries and barometer unit while the longer cylinder contains the thermometer unit. The modulated signal emitted by the meteorograph is received on a sensitive short-wave receiver, the two modulation frequencies being measured by using two audio-frequency oscillators, the output from which can be inserted in series with the receiver telephones. The frequencies of the measuring oscillators are synchronized with the incoming modulation frequencies by adjusting them until the beat frequency is zero. One of the advantages of this system is that continuous observation of pressure and temperature during an ascent is possible.

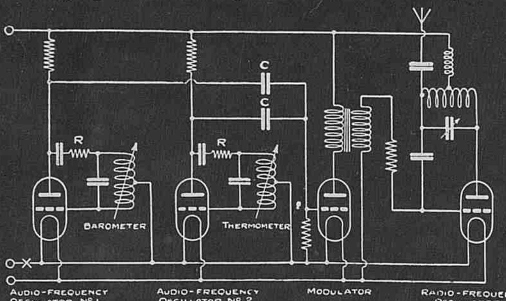
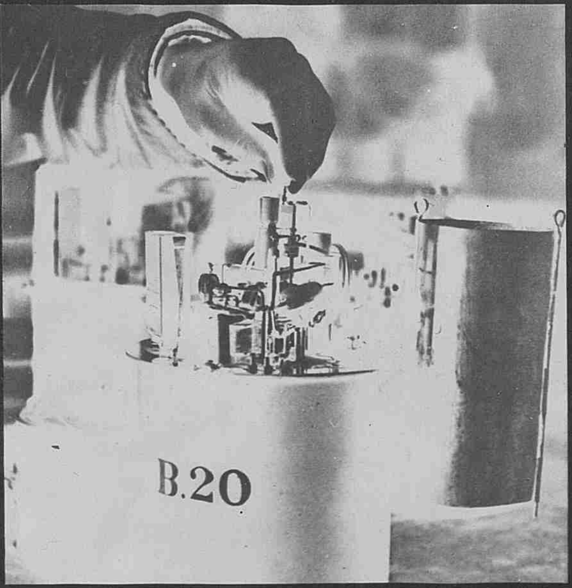

C
C
R
R
BAROMETER
THERMOMETER
AUDIO-FREQUENCY OSCILLATOR No 1
AUDIO-FREQUENCY OSCILLATOR No 2
MODULATOR
RADIO-FREQUENCY OSCILLATOR

Fig. 10. Circuit diagram of the N.P.L. radio-meteorograph (*Reproduced from the 'Proceedings of the Royal Society'*)



London News Agency Photos, Ltd.

Fig. 11. The N.P.L. radio-meteorograph

The radio-meteorograph developed in the United States by Diamond, Hinman, Dunmore and Lapham (1940) is also of the type in which the meteorological units are arranged to control the audio-frequency modulation of the carrier wave of a transmitter. In this case, however, the control is effected by variation of the resistance of the radio circuit which is shown in Fig. 12. The transmitter employs a double triode valve, one of the triodes being used as the high-frequency carrier oscillator with the variable resistance-capacity network in its grid circuit. A photograph of the radio-meteorograph is reproduced in Fig. 13. The aneroid barometer unit *c* moves a switch arm over a set of contacts, certain of which are arranged to cause the emission of signals of predetermined audio frequency. These contacts are so spaced that for a known change in pressure the arm will move from one contact to the next thus providing, by the sequence of switching operations, an indication of the pressure levels during an ascent. The other contacts are connected to a resistance which is controlled by a hygrometer *g*. In the earlier design of the apparatus the hygrometer was of the hair type, but in the latest pattern an electrical hygrometer developed by Dunmore (1939) is employed. This makes use of the variation with humidity of the electrical resistance formed by the leakage path across a thin film containing a hygroscopic salt. The actual hygrometer element consists of a bifilar winding of wire on a glass tube treated with lithium chloride. Temperature indications are obtained when the barometer switch arm is passing over insulated strips between the contacts, the frequency then being determined by the resistance of a thermometer element *f* consisting of a glass capillary tube filled with an electrolyte which has a high temperature coefficient of electrical resistance. Thus in each sequence of switching operations the pressure level is indicated by actual count of signals, while the temperature and humidity are obtained by measuring the modulation frequency of the carrier wave. The transmission is received on a short-wave receiver and is fed into a frequency meter which measures the frequency of the modulating signal and converts it into a direct current proportional to the frequency. This is recorded on a recording micro-ammeter.

Before leaving the subject of radio-meteorographs it should be mentioned that they are used not only for measurement of temperature and humidity but also for the measurement of wind velocity and direction in the upper air. This is done by employing radio direction finding methods for tracking the path of the balloon transmitter and then deriving the velocity and direction from the bearings in much the same way as when visual observations of pilot balloons with theodolites are used. A loop direction finder for ultra-short waves, which has been specially designed for use with balloon transmitters, is described by Hopkins (1938). The great advantage of the radio method over the visual method for such work is that the observations can still be made while the sounding balloon is in or above clouds.

CONCLUSION

It is perhaps appropriate to conclude this article by referring briefly to some further developments of meteorological instruments that are considered desirable.

In the case of anemometers the greatest need is for an inexpensive instrument which will indicate the instantaneous wind direction and velocity and heavy gusts directly and also give the average direction and velocity, or integrated run of wind, over a period of, say, 10 min. The instrument should also be capable of indication at a moderate distance.

For the recording of relative humidity some better method than the hair hygrograph is highly desirable, for this instrument is not capable of any great accuracy nor of operating effectively at low temperatures. Electric hygrometers, in which the surface leakage over a hygroscopic film is measured, have found some application in radio-meteorographs but they have not yet replaced the hair hygrograph for continuous recording over long periods, possibly because they are subject to large temperature effects.

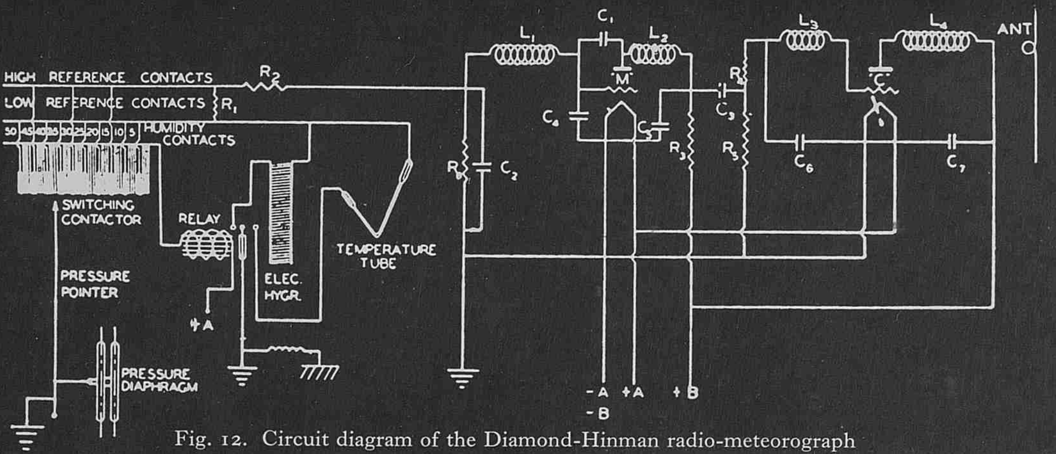


Fig. 12. Circuit diagram of the Diamond-Hinman radio-meteorograph
(Reproduced from the 'Journal of Research of the National Bureau of Standards')

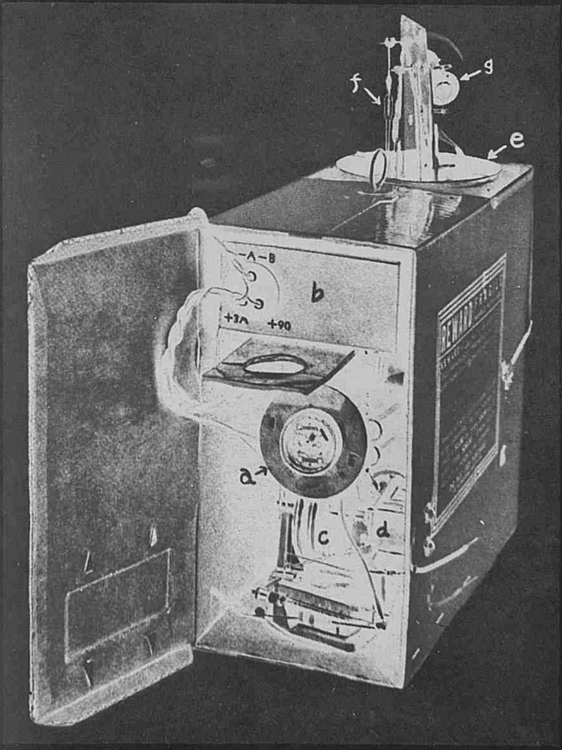


Fig. 13. The Diamond-Hinman radio-meteorograph
(Reproduced from the 'Journal of Research of the National Bureau of Standards')

The problem of measuring cloud height is one for which there is a simple solution at night, but attempts to employ a similar method in day time involve much more elaborate apparatus. The method which has been widely adopted for measurements at night is that in which the angular elevation of the spot of light produced on the base of a cloud by the vertically directed beam of a searchlight is measured by means of an alidade at a known distance from the searchlight. In order to extend this method for use in daytime it is necessary to employ a modulated or an impulsive light source and a photoelectric receiver sufficiently sensitive to detect the light spot, which is, so to speak, labelled by its modulation or impulse, against the general daylight illumination.

Finally, the question of the design of instruments for meteorological observations in modern aircraft is one that continues to be of importance despite the fact that the use of radio-meteorographs is increasing rapidly. Probably the most difficult problem here is to obtain a satisfactory method of measuring humidity at altitudes where the temperature is well below freezing point and where the normal methods of measurement by means of a wet and dry bulb thermometer or by the hair hygrometer are unsatisfactory.

It is hoped that this brief survey will give some idea of the kind of problems involved in instrumental meteorology and that instrument designers, both amateur and professional, who find these problems of sufficient interest to attempt their solution will offer their suggestions to the Instruments Branch of the Meteorological Office where they will be carefully considered.

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AIR MINISTRY

MRP 156 (Revised)
7th April, 1944.

METEOROLOGICAL RESEARCH COMMITTEE

Note on the Errors of Cup Anemometers in Fluctuating Winds.

by F.J.Scrase and P.A.Sheppard.

[This note includes comments made on M.R.P.156 by P.A.Sheppard, M.R.P.162]

1. The purpose of this note is to place on record some information which has been obtained on the extent to which cup anemometers over-estimate the mean speed of a gusty wind and on the differences, in this respect, shown by conical and hemispherical cups. We are indebted to the Meteorological Section, Ministry of Supply, Porton for the help given in carrying out the wind tunnel tests described below.

2. Some theoretical data on the question of over-estimation were included in the published description of the Sheppard anemometer (J.Sci.Inst.17, 1940, 218). The data were computed from theoretical results developed by O. Schrenk (Zeit. Techn. Phys.10, 1937, 57). In the equation of motion of a cup system

$$\frac{I}{R}\frac{du}{dt} = M - F$$

Schrenk assumes the following approximate expression for the aerodynamic couple M:-

$$\frac{M}{\frac{1}{2}\rho u^2 r^2 \pi R^2} = \lambda\left(\frac{v}{u}\right)^2 + \mu\left(\frac{v}{u}\right) + \nu$$

where I is the moment of inertia of the cup system, R the cup arm length, u the circumferential cup speed, F the frictional moment, ρ air density, r the cup radius (for hemispherical cups), v the mean wind speed and λ, μ and ν are constants determined from measurements of the torque required to turn the anemometer at various speeds in a steady air stream. By integration of the equation and consideration of extreme cases, Schrenk derived a family of curves which, to a first approximation, shows the dependence of the inertia errors on the gustiness (expressed by the ratio of the amplitude of the velocity fluctuations to the mean wind speed $\Delta v/v$,) and on the dimensionless quantity K given by

$$K = 0.55\,\rho R^2 r^2 T v / I,$$

where T is the period of the velocity fluctuations (the technical system of units, m.,kg.,sec, are used in this expression). Schrenk's curves are reproduced in Fig. 1, the left-hand vertical scale giving errors expressed as a percentage of the mean velocity of sinusoidal fluctuations and that on the right giving the errors expected for the irregular fluctuations that occur in practice. The extreme cases occur (a) when K is zero i.e. either when the anemometer has infinite inertia and (b) when K is infinite, i.e. when the anemometer has no inertia.

3. Owing to a numerical error the computed data given for the percentage over-estimation of the Sheppard anemometer in the paper referred to above are incorrect. The correct values are given in the table below together with corresponding values for the standard Meteorological Office pattern cup anemometer, the dimensions of the anemometers being:-

...

	Sheppard Anemometer	M.O.pattern anemometer.	
R	0.07	0.16	metres
r	0.025	0.063	metres
I	1.1×10^{-5}	165×10^{-5}	$kg.m.sec^2$
ρ	$0.125\ kg.sec^2/m^4$		

Over-estimation of Sheppard and M.O.Anemometers computed from Schrenk's analysis.

v (m/sec)	2			5			10		
T (sec)	1	10	60	1	10	60	1	10	60
$\Delta v/v$	0.2	0.4	0.6	0.2	0.4	0.6	0.2	0.4	0.6
K for Sheppard	.038	.38	2.28	.09	.95	5.7	.19	1.9	11.4
K for M.O.pattern	.009	.09	.52	.02	.21	1.29	.04	.43	2.58
Over-estimation (% of v) for Sheppard	1.4	3.9	1.8	1.4	2.3	0.6	1.3	1.2	0.1
for M.O.pattern	1.4	5.7	7.8	1.4	5.0	4.0	1.4	3.8	1.5

It will be seen that these figures indicate practically no difference between the errors of the two anemometers for small fluctuations of short period and that the Sheppard instrument is less subject to over-estimation than the M.O.pattern for large and long period fluctuations.

4. In order to obtain some experimental evidence of the extent of the over-estimation some comparisons of Sheppard and M.O. pattern anemometers in steady and fluctuating winds were arranged to be carried out in the wind tunnel of the Meteorological Section, Porton. According to these measurements the Sheppard anemometer showed considerably less over-estimation than Schrenk's theory indicated. Thus for a wind speed of 4.75 m./sec., T = 1.5 sec. and $\Delta v/v = 0.36$, the observed over-estimations averaged 0.5% (actual errors ranged from -1.5 to +2.5%), whereas the computed value was 4.5%. The standard M.O.pattern anemometer, on the other hand, gave results more in accordance with Schrenk's theory; for a wind speed of 4.0 m./sec., T = 1.0 sec. $\Delta v/v = 0.25$, the observed over-estimation was about 3.5% as compared with 2.5% computed.

5. It was suggested by the Meteorological Section, Porton that the closer agreement of the M.O.anemometer with Schrenk's theory was accounted for by the fact that Schrenk used a hemispherical cup form similar to that of the M.O.anemometer whereas the Sheppard instrument has a conical cup form. Some further experiments were therefore made with an M.O. anemometer fitted with conical cups of exactly the same form as those of the Sheppard instrument but having the larger dimensions of the standard cups. It was then found that there was practically no difference between the M.O. anemometer and the Sheppard instrument under approximately the same conditions as before, i.e. short period fluctuations of wind.

6. On account of the gross empiricism of the aerodynamics of the cup anemometer one can only expect Schenk's analysis to be formally correct; the numerical constants in his work are likely to vary as soon as any parameter whatever of the cup system is changed, e.g. cup form, ratio r/R, beaded or unbeaded edge. As far as the design of the cup anemometer is concerned, Schenk's quantity K may be written,

$$K \propto R^2 r^2/I$$

(If the cup is not hemispherical in form r may be defined as an equivalent radius of the cup and will be of the same order of size as the radius of the cup face).

Thus,

$$K \propto \frac{R^2 r^2}{M(G^2+R^2)}$$

where G = radius of gyration of cup about a diametral axis,

$$\propto \frac{R^2 r^2}{2\pi r^2 \sigma t(\frac{2}{3}r^2+R^2)}$$

σ = density of cup material,

t = thickness of cup material,

$$\propto \frac{1}{\sigma t(\frac{2}{3}\frac{r^2}{R^2}+1)}$$

Since the ratio r/R is appreciably less than unity, K, and therefore the extent of over-estimation, is practically independent of this geometrical factor and of r and R separately. In a given cup it depends mainly on the product σt, i.e. the mass per unit area of the cup. For the Sheppard and the M.O.pattern anemometers the mass per unit area of the cups is approximately 0.15 and 0.59 gm./cm^2 respectively. Reference to Fig.1 and Schrenk's expression for K (page 1) suggests that this difference might cause appreciable difference in the amount of over-estimation of wind speed for the intermediate range of K (intermediate values of the product $T\sigma$) and relatively large fluctuation ratios.

7. Our general conclusions are, firstly, that although Schrenk's theory gives reasonably good values for the over-estimation errors of hemispherical cup anemometers it gives excessive values for conical cup instruments (this, however, is no criticism of Schrenk's treatment), secondly, that conical cup anemometers over-estimate the mean speed of a gusty wind to a much less extent than do hemispherical cup instruments. It was not practicable to extend the experimental comparisons to high wind speeds but since the driving force increases with wind velocity more rapidly then the inertia forces, the over-estimation should become relatively less important at high speeds - as is indeed indicated by Schrenk's expression for K and fig. 1.

8. Some earlier comparisons of hemispherical and conical cups in steady winds had shown no particular advantage in the conical form apart from a tendency for the conical cups to start moving at slightly lower speeds. The difference was so slight, however, that at the time it was not thought worth while changing from the hemispherical form in the case of the M.O.pattern anemometer. In view of the results discussed above, however, it has been decided to adopt conical cups, similar in form to those of the Sheppard instrument, as standard for M.O. cup anemometers.

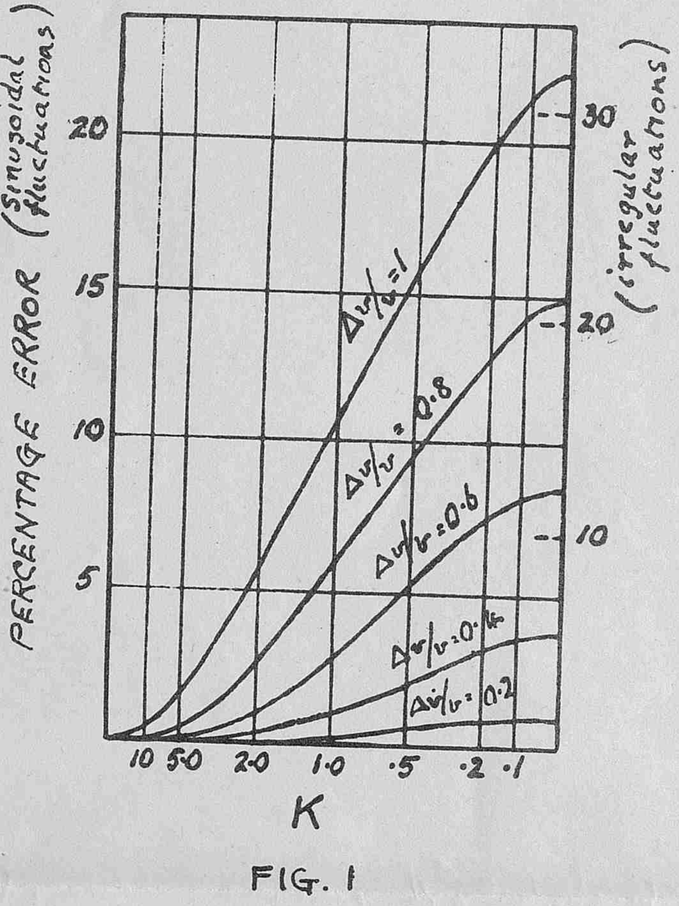
PERCENTAGE ERROR (sinusoidal fluctuations)
20
15
10
5
(irregular fluctuations)
30
20
10
Δv/v = 1
Δv/v = 0·8
Δv/v = 0·6
Δv/v = 0·4
Δv/v = 0·2
10
5·0
2·0
1·0
·5
·2
·1
K

FIG. 1

CORRECTION

M.R.P.163. § 3 on p.2 should commence - "In M.R.P.141 it was shown that on the average in cyclones the isopycnic surfaces are horizontal from 2 to about 8 km., i.e. the horizontal gradients of temperature and pressure in these levels are connected approximately etc."

The latter half of the third paragraph of § 4 on p.2 should read - "Curve A indicates that there is much less dense air, for the same horizontal level, towards the centre of the coming cyclone - particularly in the region from 3,000 to 15,000 feet, i.e. towards the centre of the cyclone the isopycnic surfaces are depressed."

Page.2. 4th and 3rd lines from end:- delete "in relation to the isobaric surfaces".

In 5th line on p.3 delete "isobaric".

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Two items of knowledge acquired shortly before that time, namely, that the troposphere on the average is colder and the tropopause lower in cyclones than in anticyclones, appeared, in the view of Shaw and Dines to make the convectional theory "utterly untenable". Rayleigh in 1917 encouraged Aitken to further experiment suggesting that it was a case where careful physical experiment and observation might accomplish more than calculation. Aitken's last paper was devoted to an attempt to show that "the penny and the pin" forms of revolving fluid were not essentially distinct types of motion. On the whole the feeling of meteorologists at that time seems to have been against Aitken's view, whilst mathematicians and physicists were left puzzled.

In later years meteorologists came to recognize that the average statistical results of W.H.Dines were influenced largely by the conditions existing in old cyclones which are relatively cold when they cross the British Isles. The present writer in the above-mentioned Memoir No.72 concluded that depressions were essentially vortices of the type which would be produced on a rotating earth* by evacuation of air from a central area, and that they in fact showed many points of analogy with Aitken's models, always keeping in mind that in the depressions the centrifugal force becomes of relatively small importance and that the air movement is not a balance between pressure gradient and centrifugal force, with molecular friction added at the lower bounding surface, but rather between pressure gradient and the deflecting force due to the earth's rotation, with the larger scale eddy turbulence as a connecting link between the upper air and the frictional effects of the earth's surface. (See Appendix A.)

2. Summary. From an analysis of the radio wind data now available it is now possible to speak with much more assurance on the subject of the part open to be played by convection; in fact to establish, as Aitken, 25 years ago, was unable to establish, that initially the air towards the centre of the active cyclone is

/definitely

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* An extension of Rayleigh's mathematical theory to the rotating earth was given by Brunt in 1921.

AIR MINISTRY

M.R.P.163
8th January, 1944.

METEOROLOGICAL RESEARCH COMMITTEE

Radio Wind Observations applied to the Discussion of The Theory of Depressions.

by

A.H.R.GOLDIE.

1. Introduction. Between 1915 and 1918, a controversy on the dynamics of cyclones went on in which the chief contributions were made by Aitken, W.H.Dines, Rayleigh and Shaw. A summary of these is given in the Introduction to Geophysical Memoir, No.72. Briefly it may be said that the discussions arose from Aitken's thesis that his model cyclones, produced by the upward convection of warmed air, showed close analogies with the depressions of middle latitudes.

Two items of knowledge acquired shortly before that time, namely, that the troposphere on the average is colder and the tropopause lower in cyclones than in anticyclones, appeared, in the view of Shaw and Dines to make the convectional theory "utterly untenable". Rayleigh in 1917 encouraged Aitken to further experiment suggesting that it was a case where careful physical experiment and observation might accomplish more than calculation. Aitken's last paper was devoted to an attempt to show that "the penny and the pin" forms of revolving fluid were not essentially distinct types of motion. On the whole the feeling of meteorologists at that time seems to have been against Aitken's view, whilst mathematicians and physicists were left puzzled.

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* An extension of Rayleigh's mathematical theory to the rotating earth was given by Brunt in 1921.

definitely of lower density than that in the outer environment of the cyclone; and further it is possible to indicate something of the life history of the vortex, to show that characteristically it develops as a whirl of maximum momentum in the lower levels of the troposphere with a pulling down of the tropopause, and that later the momentum is transferred, by the rising air, to the upper half of the troposphere, with raising of the tropopause; further that (probably) the depression dies as a result of adiabatic cooling. A new explanation of some of the ozone changes in depressions is suggested.

see correction.

3. Basis of Discussion. In MRP 141 it was shown that on the average in cyclones the isobaric surfaces are also isothermal surfaces and that the horizontal gradients of temperature and pressure from 2 to about 8 km. are connected approximately by the relation $\frac{\Delta\theta}{\theta} = \frac{\delta p}{p}$. That average condition, which with approximate geostrophic balance involves constancy of air mass transport (ρV) with height, can conveniently be taken as a pattern against which to compare individual stages in the life of depressions.

4. Averages of cases in January 1943. In examining certain curves of mean monthly variation of air mass transport with height it was seen that those referring to the Lerwick ascents of October and December 1942, January, February and March 1943 had distinctive characteristics. In these months there was a maximum air mass transport at a relatively low level - under 5,000 feet - and a continuous fall with height, except in some cases where there was a partial recovery in the upper troposphere. Further inquiry suggested that this was related to the weather of these months which in the north was unsettled with rather steep barometric gradients.

With a view to examining the bearing of this on depressions the ascents made at Lerwick on disturbed days (wind exceeding 50 km. per hr. at the 950 mb. level) of January 1943 were extracted and separated into two groups, - (A) those ascents made in front of a depression where pressure was falling and (B) those made in the rear part where pressure was rising.

There were 23 ascents in category A and 7 in category B. The ascents in each group had very distinct common characteristics; the average curves for each group are given in Fig.1. Curve A shows high momentum at low levels and a continuous fall of air mass transport with height from about 3,000 feet; the fall is very steep from 3,000 to about 15,000 feet. The mean directions are entered on the curves. Curve A interpreted in terms of isobaric surfaces indicates that there is less dense air, even isobaric level for level, and thus much less for the same horizontal level, towards the centre of the coming cyclone - particularly in the region from 3,000 to 15,000 feet, i.e. towards the centre of the cyclone the isopycnic surfaces are depressed in relation to the isobaric.

Curve B (ascents after depressions have passed) shows nearly the same conditions as existing towards the centre of the depression in the region from 3,000 to about 15,000 feet, but above this height the turnover in the conditions is extremely interesting - i.e. the air in the centre of the cyclone has now become more dense, the isopycnic surfaces are bulged up in relation to the isobaric surfaces. Considerable adiabatic cooling in these levels would in fact be an inevitable consequence of the continued ascent of the air.

Thus, far from the Rayleigh-Aitken thesis of lower density in the centre of active cyclones being "utterly untenable" it appears, on the basis of the radio wind observations, that we must accept it as characteristic that the air towards the centre starts initially by being of lower density in relation to its ~~isobaric~~ level and ends - perhaps as a result of adiabatic cooling - by having a higher density; further it looks as if in the absence of occlusion (though this remains to be established) the change over from lower to higher density spreads downwards from the upper troposphere as the cyclone ages.

The reasoning above is based on the assumption of at least rough geostrophic balance; Appendix I of Geoph.Mem. No.72 which deals mathematically and empirically with the motion of the air in relation to barometric gradient in the forward area of an advancing depression, shows that the departure from geostrophic balance, even at surface levels, is never great enough to affect the above conclusions; still less could the departures at higher levels affect these conclusions. In addition the above data are averages for a considerable number of cases and it seems scarcely conceivable that geostrophic departures could exist for days at a time in such a way as to give a considerable average departure in one direction.

It may be asked why the existence of the air of lower density towards the centre of an active depression has not been established by direct temperature observations. The answer is that large depressions, which are mainly phenomena of the six winter months, never begin their development over the land areas; they are essentially phenomena of the ocean and in the absence of upper air data from the Atlantic it is only by computation based on variation of wind with height that we can establish the relative density of the air in the active or developing depression.

Some cases of individual depressions are explored in the diagrams which follow.

5. Depression of 15th - 16th March 1943. Fig.2 shows the sequence of events at Lerwick during 15th - 16th March 1943 when a depression passed on an E.N.E. course, the centre, at its nearest point, being about 450 km. to northward of Lerwick. The depression at first deepened and then later fell off in intensity, the pressures along the lowest isobar on different charts being

15th				16th			
1	7	13	18h	1	7	13	18h.
998	984	976	976	984	984	-	996

Curve I (mean of 15d.0h. and 15d.6h.) is at the beginning of the pressure fall. Curve II(15d.19h.) shows the change over in mass transport - a diminution as compared with Curve I in the levels above about 500 mb. and a great increase around 800 mb; warmer air is indicated in the direction of the depression centre.

The next available ascent Curve III(16d.0h.) is in rising pressure. A great increase in speed has taken place at higher levels; the speed variation indicates denser air towards the depression centre at all levels up to 300 mb. and more particularly above 550 mb.

The ascents in this depression showed comparatively little change of wind direction with height.

6. Tropopause and Temperature changes on 15th - 16th March 1943. Fig.3 shows the upper air temperatures at Lerwick for 6h. and 19h. on 15th March. In the Upper Air Supplement the tropopause

at 19h. is described as Type II at the very low level of 400 mb. The wind direction at 19h. was not very greatly different from that at 6h. and it seems most unlikely that any difference in origin of the air could account for the difference in tropopause level and in stratosphere temperature between 6h. and 19h.

It occurs to one therefore to inquire whether this peculiar distribution of temperature and the low tropopause might be a dynamical product of the depression, in other words to examine what pressure changes impressed upon the air shown in curve A could turn it as a result of adiabatic changes into curve B. The calculation has been done for the levels from 700 to 150 mb. with the following results:-

Pressure level (mb)	Pressure change required (mb)
150	+9
200	+21
250	+59
300	+67
350	+42
400	+12
450	- 3
500	-20
550	-24
600	-26
700	-10

This points to upward motion in the region below the 400mb. level and downward motion in the stratosphere. A comparison of curves A and B with this in mind and using a Hertz diagram suggested the following picture:-

(a) about half the air between the initial and final tropopause levels (250 mb. and 400 mb.) evacuated (presumably by horizontal divergence);

(b) the air in the troposphere ascending - (by some 2,500 feet in the middle levels) and diverging under the tropopause; temperature change would mainly be along the saturated adiabatic;

(c) the air originally above about the 330 mb. level, i.e. mainly stratosphere air, subsiding and being filled in above to some extent by convergence in still higher levels, mainly above 200 mb., temperature change would be along the dry adiabatic;

(d) the high level inflow of air - equivalent to some 60 mb. - above the 200 mb. level ought to raise the ozone amount by 25 or 30% ; this and other features noted in this paper fit in with Dr.Dobson's results a, b, c and d on page 23 of MRP 37. Some further cases of depressions which have been examined give the same general picture of the changes in stratosphere and higher troposphere levels.

7. Vortical Intensity Changes on 15th - 16th March 1943.

The wind data of Fig.2 can be used further to get a rough picture of the vortical development of the depression during 30 hours. The distances of Lerwick from the isobaric centre of the depression (at surface level) at the times to which curves I, II, III, and IV relate were approximately 550, 450, 600 and 850 km. respectively. Assuming, as the Lerwick autographic records suggest, that the radius of maximum surface speed was about 550 to 600 km., that the velocity outside of this radius varied as in a simple vortex and inside varied directly as distance from the centre, we can take curves I and III as they stand and we should raise the speeds in curve II roughly by a fourth, and those in curve IV by a half to make them all a measure of vortical intensity on the same basis. The results are shown in Fig.4.

The vertical profile of the vortex thus develops from I to II by a great increase of momentum in the lower half of the troposphere and the letting down of the stratosphere to about 23,000 feet; the momentum in the lower half then falls off or rather is in part transferred by the rising air to the upper half of the troposphere as in III, the tropopause level being restored again; finally in IV the momentum has been distributed to even higher levels and the tropopause raised further, whilst a condition of approximately constant air mass transport at all levels in the troposphere has been established. The total momentum of the vortex is proportional to the area between the curves and the vertical axis, and is seen to be roughly of the same order in II, III and IV.

The above rough picture has not taken account of the displacement of balloons from their starting point or of the inclination of the axis of the depression to the vertical. It has also not taken account of the varying bearing of the place of observation from the centre of the depression (which changed gradually from 120° to about 210°) and thus of possible considerable departure of the system from symmetry. It might therefore be argued that the features of curves I, II, III and IV are rather features of wind direction or of air masses, and it would need a longer series of discussions with more elaborate analysis to answer this point satisfactorily.

It is also not easy, with only a few upper air ascents, to disentangle effects due to station position from changes in the structure or condition of the depression. To some extent however these points are answered by the case dealt with at the end of this note where a rapidly moving depression is taken and the speeds at Lerwick and Larkhill on either side of the path are computed in relation to the centre of the depression. It will be seen that the results still give essentially the same picture.

The picture has many points of analogy with Aitken's idea of the vortex open at the lower end and continually taking in new air in low levels and transferring it to higher levels. Being virtually capped at the higher end by the stratosphere the vortex must pass out the air ultimately by divergence under the tropopause and/or must raise the tropopause.

8. Development of a Secondary. The next case, Fig.5 relates to the sequence of events at Larkhill whilst a secondary depression developed rapidly off N.W.Ireland on 5th February, 1943. This differs from the case dealt with in Figs. 2-4 in that the development was very rapid, each curve shows a considerable variation of wind direction with height and there may have been appreciable departures of actual wind from geostrophic balance, though probably not enough to affect the general argument. Curve I (5d.6h.) indicates much less dense air, level for level, towards W.N.W. from Larkhill up to 300 mb. and thereafter towards north, i.e. roughly where the depression is developing. Curve II (5d.12h.) gives similar indications with, however, the beginning of the characteristic rise in momentum in low levels. Curve III (5d.20h.) gives a continuation of the changes shown beginning in Curve II, it indicates that there is warm air to N.N.W. all the way up. This stage corresponds to Curve II of Fig.2.

The next available ascent is at 6d.6h. (Curve IV). The depression is then moving away N.Eastward and pressure at Larkhill is rising slowly. The curve shows a tremendous rise since the previous evening in the momentum in all levels from surface up to 400 mb. The decrease above that level indicates denser air in high levels in the rear part of the central area of the depression. The next ascent at 6d.12h. Curve V (corresponding to III of Fig.2) shows the transfer of maximum momentum from the lower to the upper half of the troposphere. The distances of Larkhill from the isobaric centre at the times to which these curves relate were 800, 800, 750 and 950 km. respectively.

9. Filling up of a Depression. Fig.6 illustrates the later stages of a depression's life - the filling up in situ of a shallow depression centred to northward of Lerwick on 2nd - 3rd February 1943. Originally over the Atlantic this had been an intense depression but its intensity diminished as it approached.

Curve I (2d.12h.) shows a very pronounced maximum of momentum in low levels around 850 mb., the great decrease and slight backing with height indicating warm air towards the centre of the depression; maximum development has apparently been reached or even passed. The next curve II (mean of 2d.20h. and 3d.5h.) with much the same wind directions, though still indicating a decrease of speed with height, shows a distinct diminution on the previous rate of decrease up to 400 mb. i.e., the air towards the centre from 850 mb. to 400 mb. has become less warm. In the next curve III at 3d.12h. the momentum at low levels has fallen to half and there is little change of momentum with height up to 400 mb. The last curve IV at 3d.20h. shows virtual calm at low levels and actual increase of momentum with height up to 550 mb. i.e., definitely denser air towards the centre up to this level.

The sequence of changes here is in accord with a gradual adiabatic cooling of the depression centre throughout the troposphere and finally possibly occlusion by cold air in the lower half of the troposphere. At Lerwick itself (only some 300 km. from the depression centre) the mean drop in temperature in 24 hours as between the ascents of the 2nd and those of the 3rd was as follows:-

Levels	Drop in Temperature (°F)
900 - 700 mb.	3
550 mb.	3.5
400 - 300 mb.	7
250 - 150 mb.	0.5

This suggests that in the main the depression died through the centre, especially in the upper troposphere, becoming cold in relation to its environment and therefore most probably as a result of adiabatic cooling of ascending air.

This depression at this period of its life showed little or no activity above the tropopause, which throughout the period here discussed was at about the 300 mb. level.

10. Air Motion North and South of Centre of fast moving Depression.

In all cases dealt with above the observing station lies to the south or right of the path of the depression. Relatively seldom does the depression track come to the southward of British Stations.

In the case of 5th - 6th December 1942 however, the centre moved from a position on the Atlantic to the west of the Outer Hebrides, across the middle of Scotland and then over the North

Sea towards Denmark. The centre moved on a slightly curved path at velocities estimated as follows:-

5d.6h. 30 km./hr. from 290°
5d.12h.)
5d.20h.)40 km./hr. from 280°
6d.5h. 45 km./hr. from 275°

Since the speeds are fairly high it becomes necessary, if a reasonably accurate picture of the motion with regard to the centre is needed - particularly in the case of a station lying to the northward of the path - to compound the airspeeds at the different levels with the velocity of the centre reversed. [For surface winds entering a travelling vortex the matter is, however, different].

This has been done with the data for Lerwick and Larkhill one on each side of the track and the resultant speeds have been then transformed to air mass transport in order to give diagrams comparable with those given for earlier cases. During the period discussed the distance of the centre from Lerwick varied from about 400 km. at the nearest to about 700 km. at the farthest; from Larkhill from about 530 km. to 800 km. Thus without some further assumptions with regard to the dimensions of the vortex and a corresponding adjustment of speeds for different times at the two places, it is not legitimate to use the diagrams in Figs.7 and 8 as strictly comparable measures of vortex strength.

They are reproduced here rather to show that both north and south of a depression essentially the same picture is got of the vertical behaviour, namely at first the whirl has its great momentum in the lower levels, then later there is transfer of much of this momentum to higher levels of the troposphere as a prelude to final decay.

At Lerwick the lowering and subsequent rising of the tropopause were also noticeable in this case. The changes were as follows, but at Larkhill which was rather fabther from the centre the changes were small.

Time	Lerwick	Larkhill
4d. 20h.	Type I at 206 mb.	Type I at 188 mb.
5d. 5h.	Type II at 257 mb.	Type I at 225 mb.
12h.	Type II at 295 mb.	Type I at 187 mb.
20h.	Type I at 180 mb.	Type I at 179 mb.

11. Structure above the 200 mb. level.

The horizontal momentum of the air up to about the 80 mb. level has also been examined. Above the 200 mb. level it appears to be small, and to be not very greatly increased by depressions. In the group of depressions in January 1943 it was raised by the passage of the depression on the average by an amount which varied from some 38% about the 200 mb. level down to 16% at the 100 mb. level. The changes attending the depressions of 15th - 16th March and 2nd - 3rd February 1943 were even less; those of 5th - 6th February 1943 were somewhat greater up to the 100 mb. level, and then fell off sharply. The inference would seem to be that the effects at most are of secondary importance to those at lower levels.

The relationship of the stratosphere subsidence to the ozone changes referred to in para. 6, if it should prove to be justified, is however of importance. It provides an interpretation of the ozone observations which makes those of immediate value in detecting the areas in which depressions or secondaries are beginning to develop, or in verifying deductions made by other methods as to these areas.

12. Conclusion.

The conclusion of the writer so far therefore is that these results point in the same direction as the investigations of Geophysical Memoirs Nos. 72 and 79, namely in favour of regarding the vertical behaviour as the fundamental feature of the depression, and the changes, contortions and wrapping up of the fronts not as essentially governing factors, but rather as consequences of the travelling vortex.

Appendix A. Summary of Results given in Geoph. Memoirs Nos.72 and 79

Apart from the identification of analogies between depressions and the model cyclones of Aitken, the principal results given in these Memoirs were as follows:-

1. In the polar air part of the central area of a depression there is sometimes, but not always, an approximation of the average air speed at surface level to direct proportionality to distance from centre ($\frac{v}{r}$ = constant); it is not as a rule genuine "solid rotation" there being normally considerable incurvature of the air flow, and the dynamical conditions for a disc of air spinning like a solid and having a general translational motion are not satisfied.

2. In a certain region outside the central area the average surface air motion approximates to a simple vortex (vr = constant).

3. There is some evidence that the central area, which in an average depression has a radius of the order of 500 km., increases in diameter as the depression ages. Depressions of unusual violence in regard to wind are those in which the vortex motion is developed up to an unusually small radius from the centre.

4. There is an intimate relation between the barometric pressure at the centre of a depression, and the intensity of its outer vortex. Eighteen cases examined gave a correlation coefficient of -0.93 with a standard error of 0.03 for this relationship. [This is in accord with the mathematical theory that the intensity of the vortex originated by removal of "fluid" from a central area is proportional to the amount of the fluid removed.]

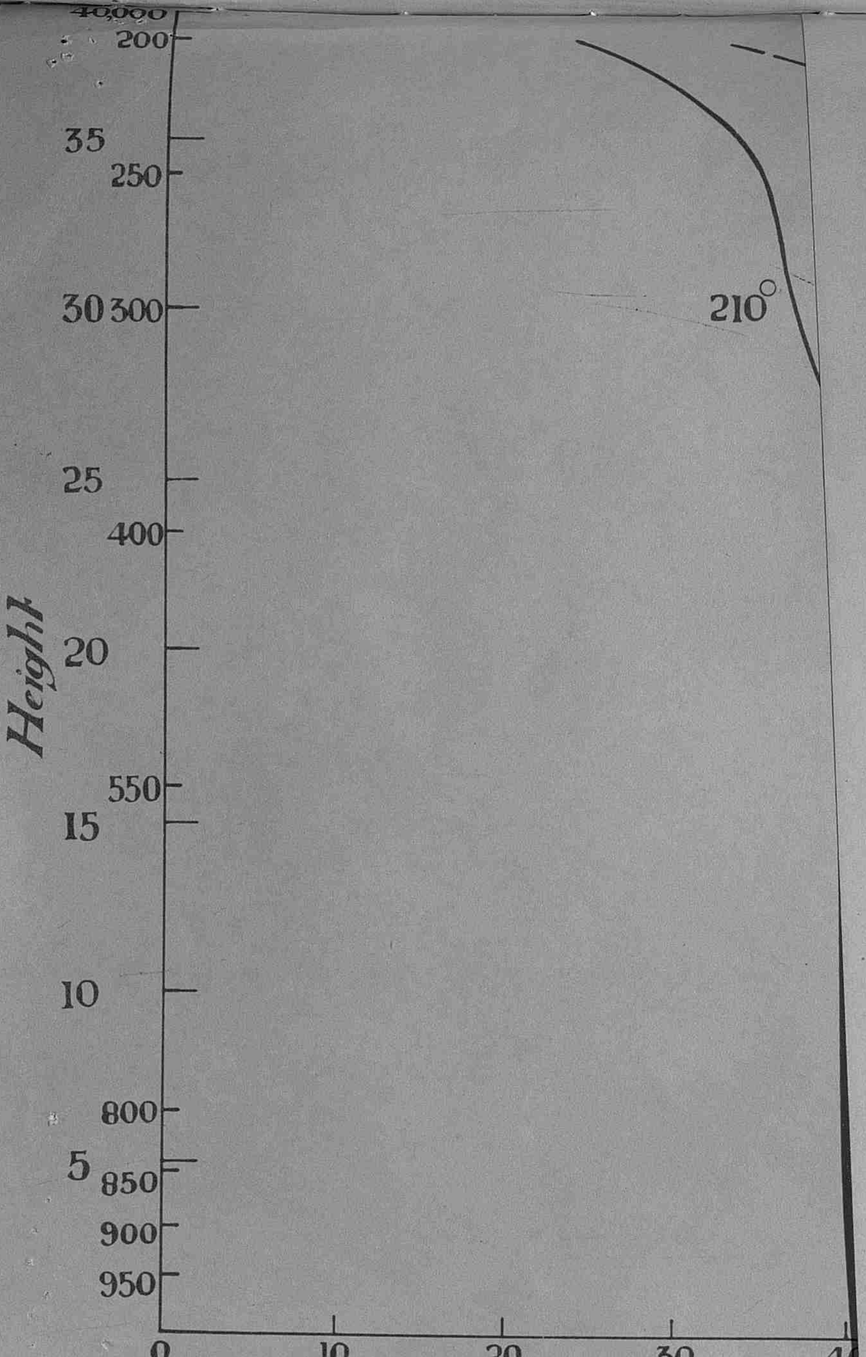
40,000
200
35
250
30
300
25
400
20
Height
550
15
10
800
5
850
900
950
0
10
20
30
4
210°
ev (

Fig. 1.

Depression of 15–16 March 1943

Variation of Momentum with height at Lerwick

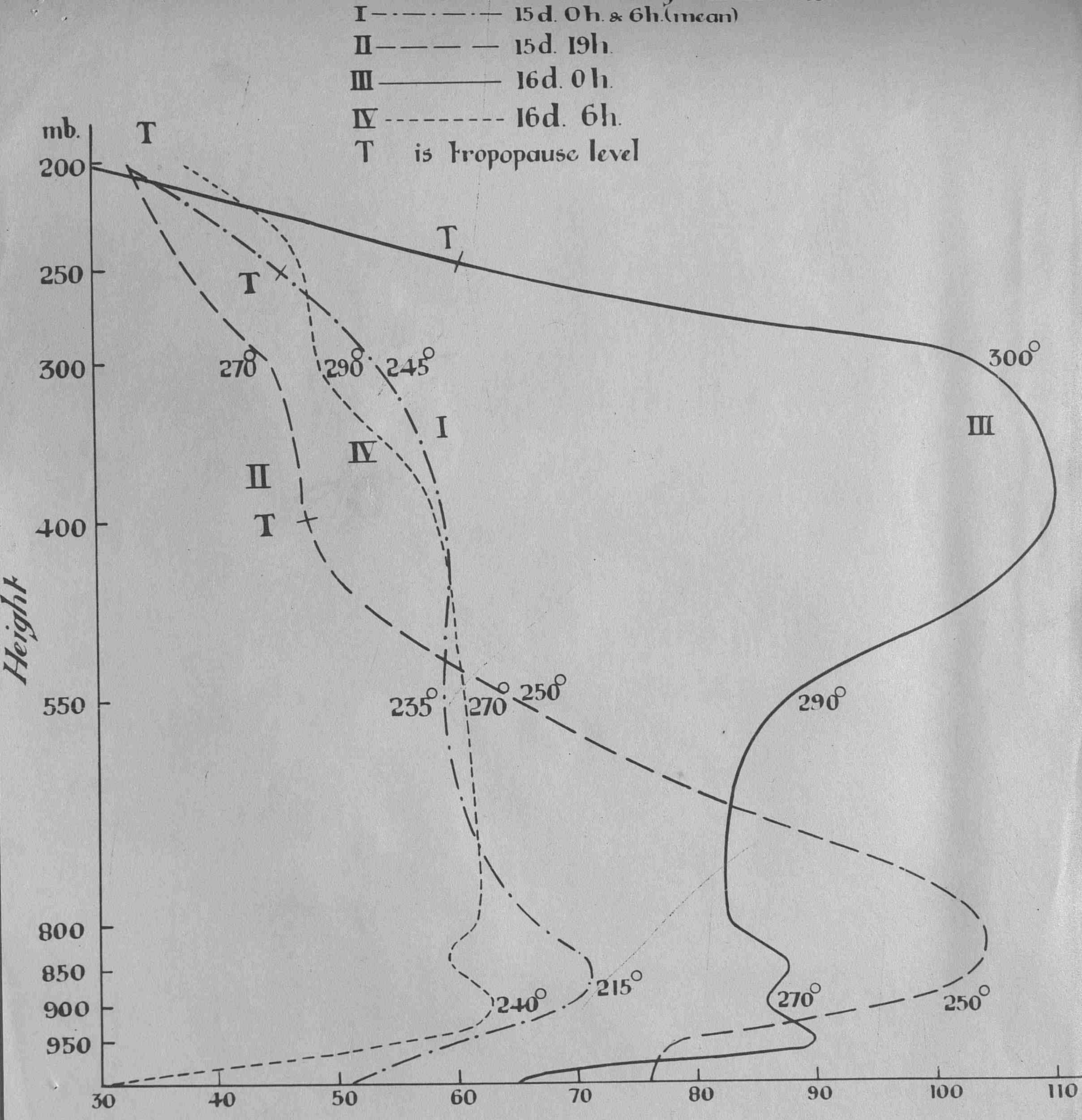

Fig.2.

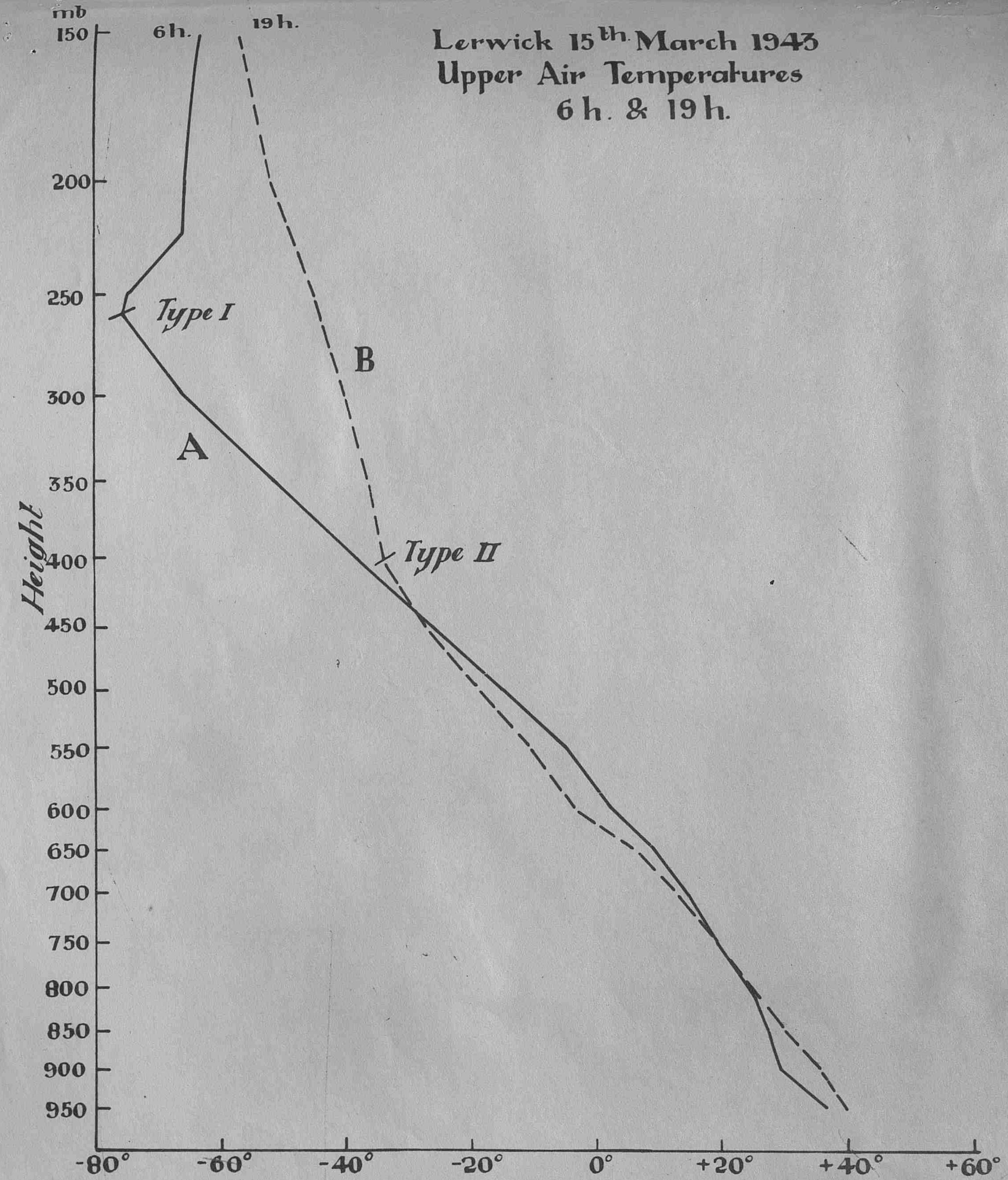
Lerwick 15th March 1943
Upper Air Temperatures
6 h. & 19 h.
mb
150
200
250
300
350
400
450
500
550
600
650
700
750
800
850
900
950
Height
6 h.
19 h.
Type I
Type II
A
B
-80°
-60°
-40°
-20°
0°
+20°
+40°
+60°
Temperature °F

Fig. 3

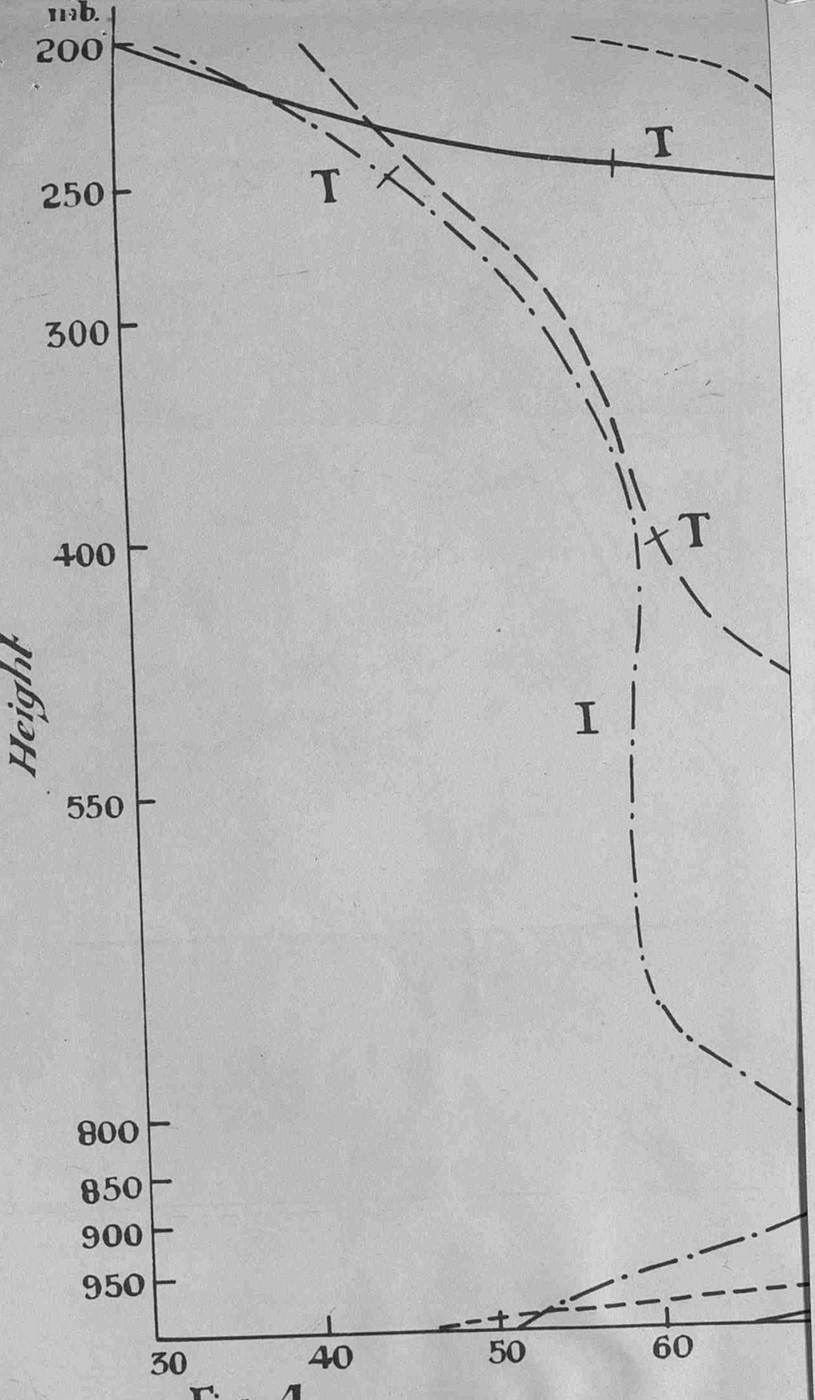

Fig. 4.

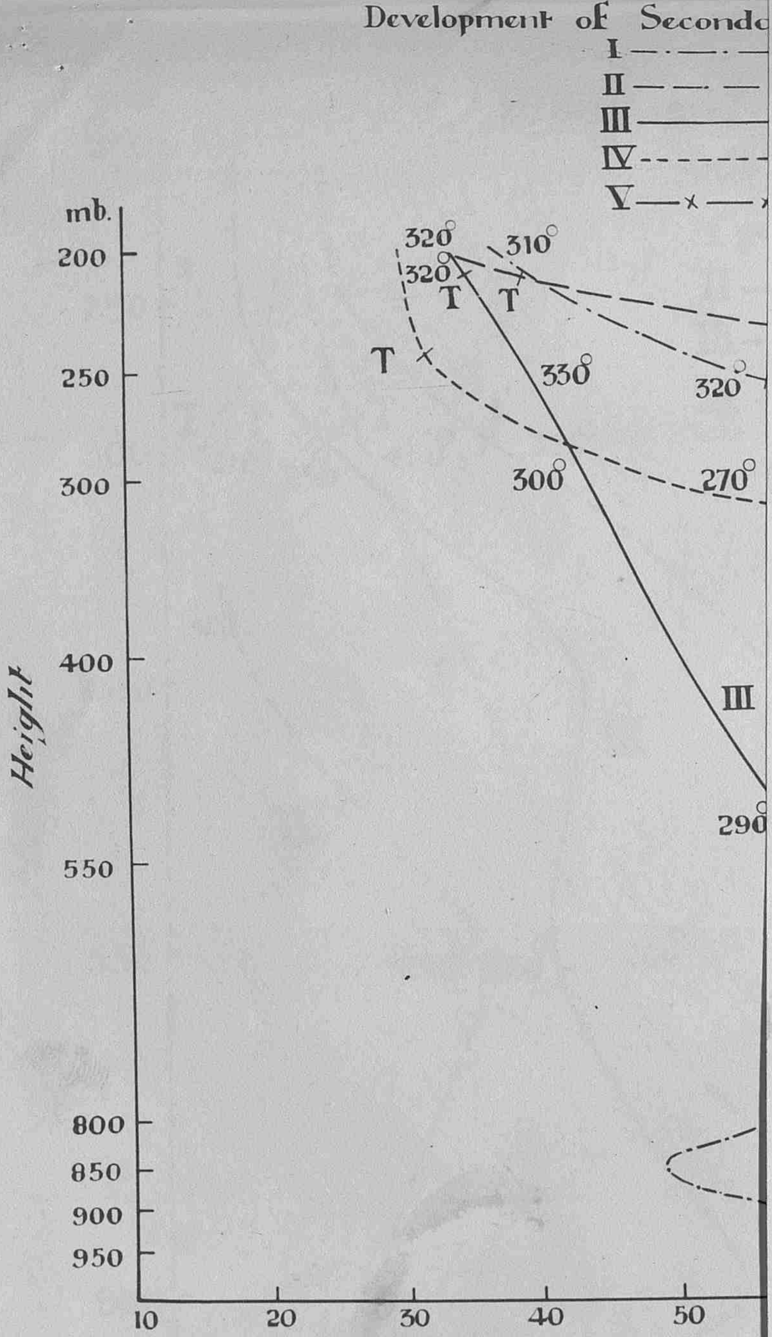
Development of Second
I
II
III
IV
V
mb.
200
250
300
400
550
800
850
900
950
Height
10
20
30
40
50
320°
320°
310°
T
T
T
330°
320°
300°
270°
III
290°

Fig. 5.

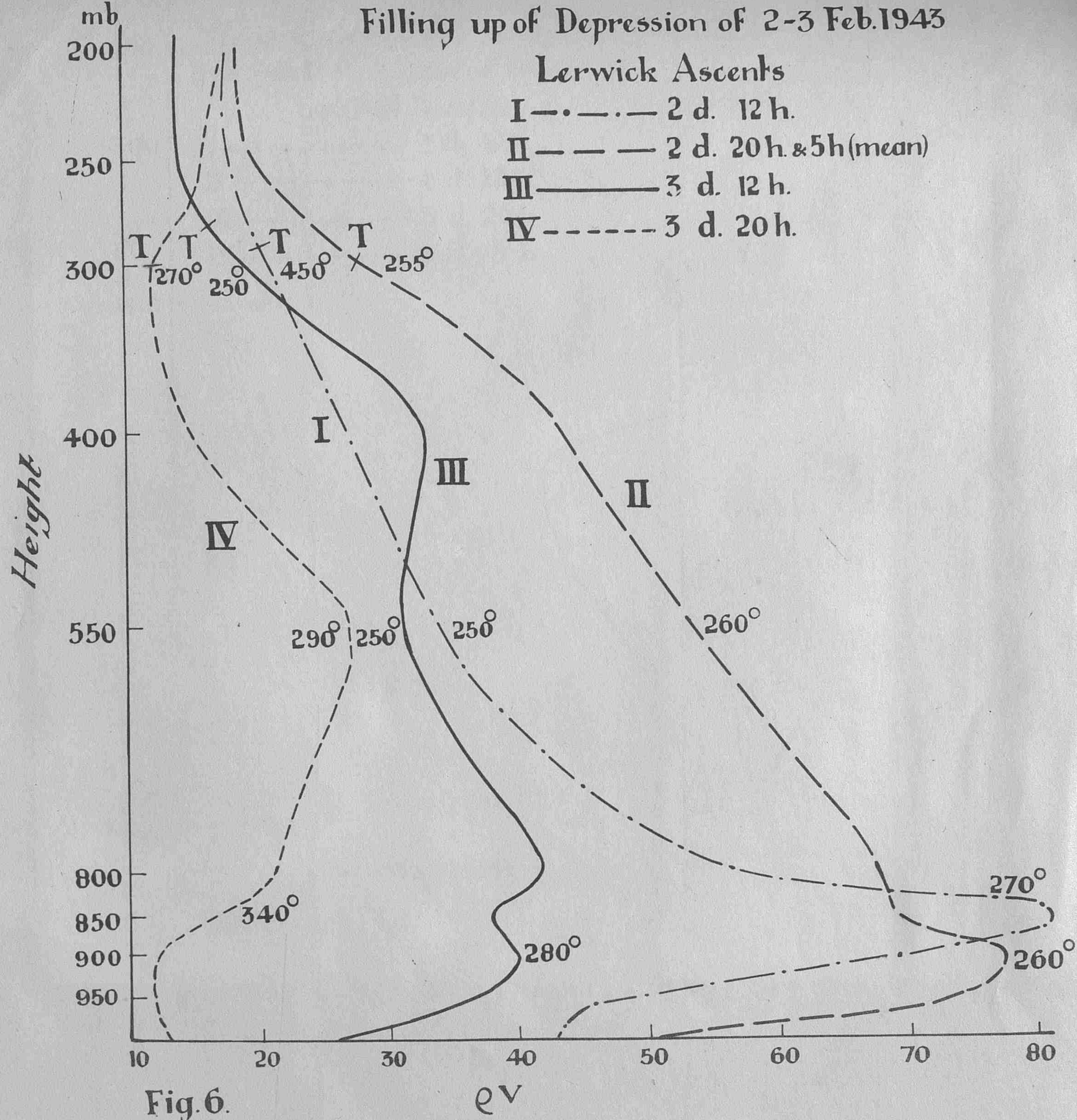
Filling up of Depression of 2-3 Feb.1943
Lerwick Ascents
I —·—·— 2 d. 12 h.
II — — — 2 d. 20 h. & 5 h (mean)
III ——— 3 d. 12 h.
IV - - - - - - 3 d. 20 h.
mb
200
250
300
400
550
800
850
900
950
Height
T
T
T
T
270°
250°
450°
255°
I
III
II
IV
290°
250°
250°
260°
340°
280°
270°
260°
10
20
30
40
50
60
70
80
ev

Fig.6.

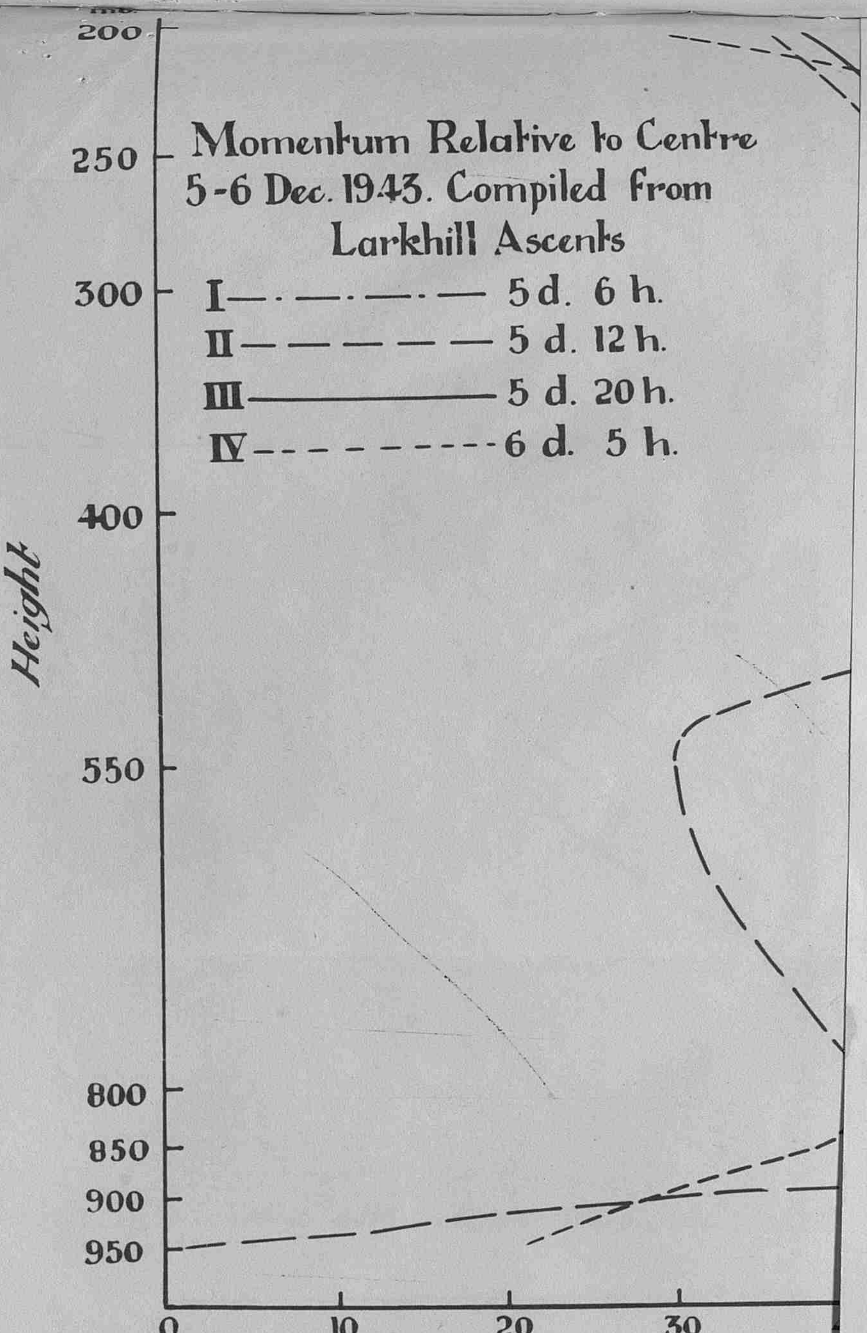
mb.
200
250
300
400
550
800
850
900
950
Height
Momentum Relative to Centre 5-6 Dec. 1943. Compiled from Larkhill Ascents
I 5 d. 6 h.
II 5 d. 12 h.
III 5 d. 20 h.
IV 6 d. 5 h.
0
10
20
30

Fig. 7.

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

M.R.P. 164

31st January, 1944.

Air movement at different levels above Lerwick and Larkhill.

by F.M.Jones.

Summary. The monthly averages of wind speed and of air mass transport at different levels at Lerwick and Larkhill are discussed. Averaged over six months or a year the air mass transport is nearly constant from just above the first kilometre up to nearly the tropopause. The monthly averages where they do not show this constancy, tend to fall into groups with characteristics which seem to be related to the prevailing pressure systems. The relation of surface to upper wind and the wind direction frequencies are referred to briefly.

A large number of radio wind measurements are now available for Lerwick and Larkhill. These have been averaged for the isobaric levels of 950, 900, 850, 800, 550, 400, 300, 250 and 200 mb., for the months April 1942 to June 1943 inclusive; the mean speeds are given in Tables I and II. The surface values are corresponding hourly means from pressure tube anemometer records.

As a rule the mean speed increases more or less considerably with height up to the level of the tropopause, but since the density diminishes, the relative flows at different levels are brought into much better perspective with one another by calculating the values of ρv - the air mass transport. The values of ρv for year, winter and summer, and for individual months are shown in Figures 1, 2 and 3. From Fig.1 it is seen that ρv is practically constant at Larkhill from 900 mb. to 300 mb. in the summer and winter means. At Lerwick there is no great departure from a similar constancy in summer, but in winter ρv reaches a maximum at about 4,000 feet; thereafter it diminishes slowly. Fig.1 shows how much greater the range between summer and winter conditions is at Lerwick than at Larkhill. The values of ρv for individual months show bigger variations; the curves which do not show (A) relative constancy with height depart from that condition in two ways - (B) by a maximum transport somewhere about the middle troposphere, giving the curve a "barrel shape" or (C) by a minimum transport either around the 850 mb. level or just under the tropopause, or at both these levels, with a minimum in the middle troposphere, thus giving the curve a (more or less deformed) "hour-glass shape".

Since the rate of increase of horizontal barometric gradient with height in mb. per metre increase of height is given by

$$\cdot 0342 \frac{p}{\theta}\left(\frac{\Delta\theta}{\theta} - \frac{\Delta p}{p}\right)$$

the A or average condition implies

$$\frac{\Delta\theta}{\theta} = \frac{\Delta p}{p}$$

i.e., the same density is found on the same ~~isobaric~~ level. The B or barrel-shaped curves mean that $\frac{\Delta\theta}{\theta} > \frac{\Delta p}{p}$ in the lower half of the troposphere, and $\frac{\Delta\theta}{\theta} < \frac{\Delta p}{p}$ in the upper half.

Investigations* made by Dr.Goldie indicate that the B type is associated mainly with high pressure areas, and thus we get the result that in the central area of highs the air in the upper half of the troposphere is relatively dense in relation to its isobaric level, and the air in the lower half is relatively warm.

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* M.R.P.163.

The C or hour-glass type of curve is associated, according to the same investigations, (M.R.P.163) with cyclones; C curves may be subdivided into two types, C1 where the lower bulge of the "hour-glass" is exaggerated, and C2 where the upper bulge is exaggerated. The interpretation of C1 by the formula quoted above is that in the central area of the cyclone the air of the troposphere is relatively light or warm for its isobaric level. The meaning of the type C2 is that the air in the centre of the low pressure, especially in the upper half of the troposphere, is relatively dense or cold for its isobaric level. Again, according to the same investigation the cyclone runs first through stage C1 and finally through C2. It is therefore to be expected that in monthly mean curves the cyclonic types C1 and C2 will predominate at any northern station in winter, whilst at a southern station, nearer to the semipermanent highs, the types B or A should be more in evidence. A comparison of the curves in Figs.2 and 3 with the wind and mean pressure maps of the Monthly Weather Report illustrates these points. Thus, September, October and December 1942, and January and February 1943, which show curves of marked C1 type for Lerwick, were all months with steep barometric gradients and relatively low pressure in the north. April 1942, which shows a B curve for Lerwick, was influenced by an anticyclone over Scandinavia. November (B at both places) was quiet and dry, with pressure exceeding the average very substantially, the excesses ranging from 9.5 mb in S.E.England to 15 mb. in the Outer Hebrides.

In view of the variety of conditions it is not altogether easy to compare the flow of air at Lerwick with that at Larkhill, or to compare one month with another. Table III and Fig.4 give the mean flow at each place for the three levels 250, 550 and 850 mb. taken together. These figures should give a fairly reliable index to the integral of the air transport with respect to height. There is a suggestion that, as the total transport goes down below the seasonal normal at one place it goes up at the other. A comparison of the wind speeds in April, May and June 1942 with those of the same months in 1943 suggests however, that a good many years' observations would be needed to arrive at reliable monthly normals, particularly for Lerwick.

The relationship of the surface (anemometer) values of ρ v to the general drift at levels high enough to be free from surface effects is of interest. This general drift is taken as the mean of ρ v at 800 mb. and 550 mb. (roughly 6,500 and 16,000 feet respectively). The percentage relationships of the surface ρ v to the upper ρ v are as follows:-

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Win.	Sum.
Lerwick	85	74	81	77	80	72	82	74	76	81	70	75	75	76
Larkhill	49	41	45	69	53	48	52	46	47	37	31	47	43	54

The strength of the surface wind in relation to the upper wind is a balance between the opposed effects of surface friction tending to bring the surface air to rest, and of turbulence tending to equalize momenta of surface air and upper levels. The constancy of the ratio at Lerwick is notable; it varies only between 85% in Jan. and 70% in Nov. At Larkhill the values are much lower than at Lerwick, presumably because of the greater surface friction of a large land area; whilst the annual range, from 69% in April to 31% in November is much greater than at Lerwick, presumably a reflection of the larger variation of turbulence over land as between spring and early winter.

Wind frequencies for the four seasons for the 250 and 550 mb levels are given in tables IV and V. The almost complete absence of winds from directions between about 30° and 160°, except in spring is noticeable.

Table I. Mean speed in Km./hr. at LERWICK.

Height in mb.	Surface	950 2000'	900 3000'	850 5000'	800 6500'	550 16000'	400 23500'	300 30000	250 34000'	200 38000
1942 April	22	27	27	28	30	50	58	58	53	42
May	26	34	35	37	38	54	68	77	70	50
June	28	35	36	39	44	65	96	115	103	76
July	23	30	31	32	34	44	57	76	76	65
August	23	33	35	37	39	50	64	75	73	59
September	31	42	48	50	52	63	83	99	102	92
October	35	42	47	50	54	68	94	118	121	106
November	35	47	50	55	58	83	105	115	112	95
December	37	50	56	60	61	77	97	119	114	95
1943 January	32	48	51	50	49	54	67	83	85	79
February	44	58	65	70	72	93	112	134	140	136
March	37	47	53	57	58	69	88	105	105	97
April	37	45	51	56	60	80	107	123	122	117
May	31	37	40	41	42	59	83	93	92	71
June	23	32	33	34	35	43	53	59	57	48

Table II. Mean speed in Km./hr. at LARKHILL.

Height in mb.	Surface	950	900	850	800	550	400	300	250	200
1942 April	25	42	46	48	49	52	63	73	70	61
May	17	31	34	36	39	52	73	91	90	76
June	10	19	21	21	20	39	54	67	61	54
July	19	31	35	38	41	65	86	98	103	93
August	15	25	30	34	38	55	68	80	84	71
September	15	30	33	35	37	53	72	89	90	74
October	15	34	40	44	48	68	87	102	103	90
November	9	29	29	30	31	52	68	82	83	75
December	21	43	48	53	56	69	91	103	102	88
1943 January	22	41	45	48	51	74	96	106	103	89
February	17	39	45	48	49	67	87	102	103	93
March	12	27	29	31	32	44	57	69	72	53
April	17	38	43	45	49	69	81	94	92	82
May	16	35	39	41	43	60	72	78	78	66
June	13	30	35	37	39	58	74	93	99	80

Table III. e v, Lerwick and Larkhill, Mean of 250 mb., 550 mb., and 850 mb., levels.

Month	Lerwick	Larkhill
1942 April	30	39
May	36	38
June	43	25
July	32	42
August	35	36
September	47	37
October	51	46
November	56	35
December	57	50
1943 January	43	50
February	68	48
March	52	32
April	57	45
May	42	40
June	30	40

Table IVa. % Frequency Wind direction, LERWICK, 250 mb.

	10-20	30-40	50-60	70-80	90-100	110-120	130-140	150-160	170-180	190-200	210-220	230-240	250-260	270-280	290-300	310-320	330-340	350-360
Spring	4	2	1	3	3	6	6	3	3	6	7	7	14	11	14	3	2	4
Summer	3	1	1	0		1	0	3	4	7	7	10	10	16	11	10	7	7
Autumn	1	1	0	0	1		1	2	2	8	9	13	10	17	12	11	6	5
Winter			0		0		1	3	3	7	9	12	13	14	15	13	6	1

Table IVb. % Frequency wind direction, LERWICK, 550mb.

	10-20	30-40	50-60	70-80	90-100	110-120	130-140	150-160	170-180	190-200	210-220	230-240	250-260	270-280	290-300	310-320	330-340	350-360
Spring	3	0	2	1	3	7	10	8	3	4	7	7	14	9	7	5	1	7
Summer	2		0	0	1	1	0	3	6	6	8	11	12	12	19	7	5	6
Autumn	1	0	0		1	1	1	1	1	12	10	9	15	14	11	11	6	4
Winter	1	1		0			0	5	3	9	11	10	10	17	13	10	6	1

Table Va. % Frequency wind direction, LARKHILL, 250 mb.

	10-20	30-40	50-60	70-80	90-100	110-120	130-140	150-160	170-180	190-200	210-220	230-240	250-260	270-280	290-300	310-320	330-340	350-360
Spring	6	3	4	4	3	4	2	4	2	4	11	9	11	7	8	8	4	4
Summer	5	1	0		1	0	1	2	4	7	5	13	14	12	11	10	9	4
Autumn	4	3	3	1	0	1	1	1	3	4	10	9	.9	10	9	10	12	8
Winter	3	2		0	1	0	0	1	1	7	12	10	12	9	8	17	11	3

Table Vb. % Frequency wind direction, LARKHILL, 550mb.

	10-20	30-40	50-60	70-80	90-100	110-120	130-140	150-160	170-180	190-200	210-220	230-240	250-260	270-280	290-300	310-320	330-340	350-360
Spring	3	4	2	5	5	4	5	5	5	5	8	11	10	6	9	7	2	4
Summer	3	1	0			0	1	1	2	9	6	12	15	10	13	13	7	4
Autumn	3	3	2	1	1	0	1	1	2	6	10	13	13	13	10	10	7	4
Winter	3		2	0		0	1		3	5	19	11	9	8	11	16	7	4

Fig. 1. Variation of Air Mass Transport with Height

(Seasonal means)

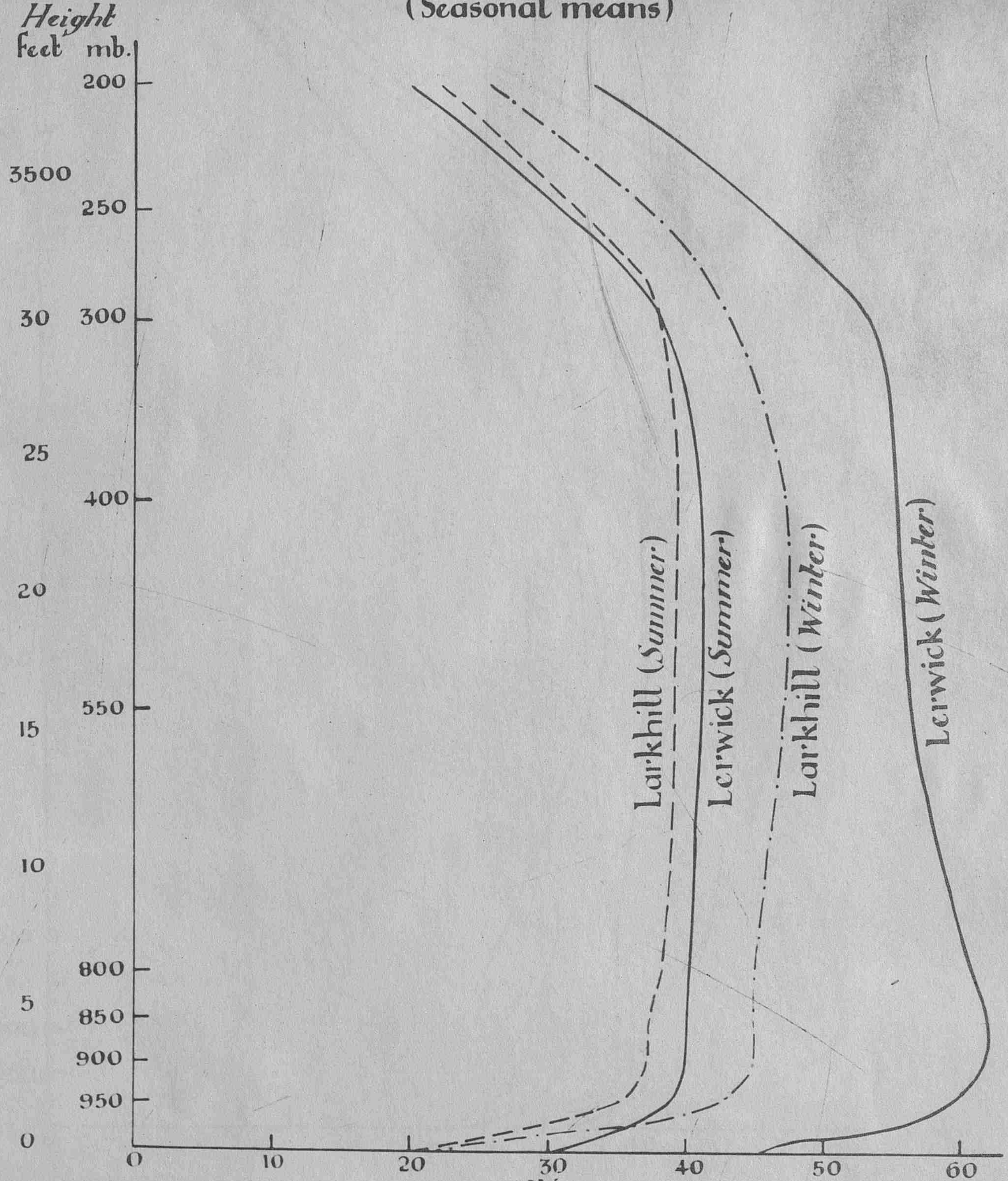


Fig. 2a. Variation of Air Mass Transport with Height. (Monthly means) Lerwick, Summer.

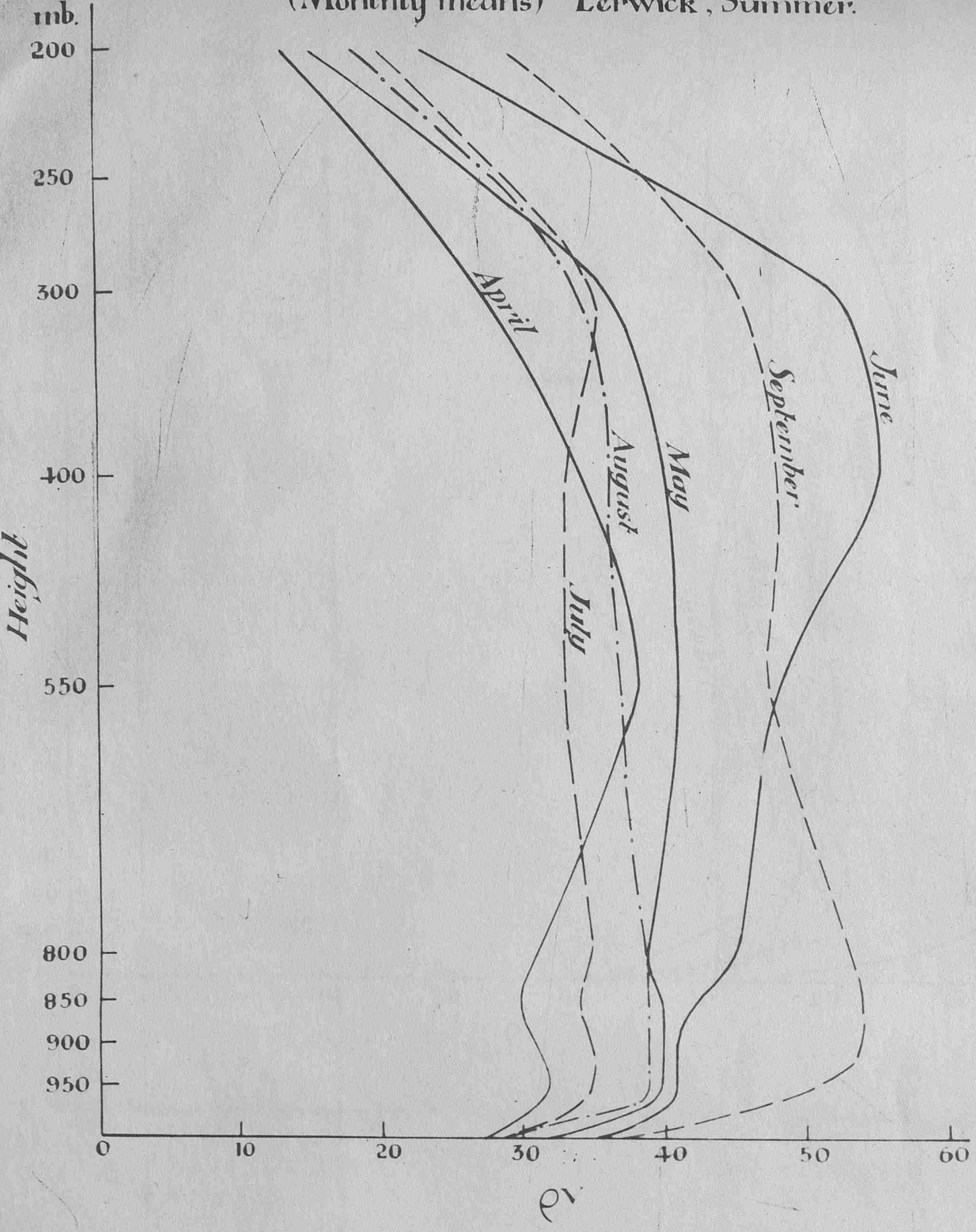


Fig. 2b. Variation of Air Mass Transport with Height (Monthly means) Lerwick, Winter.

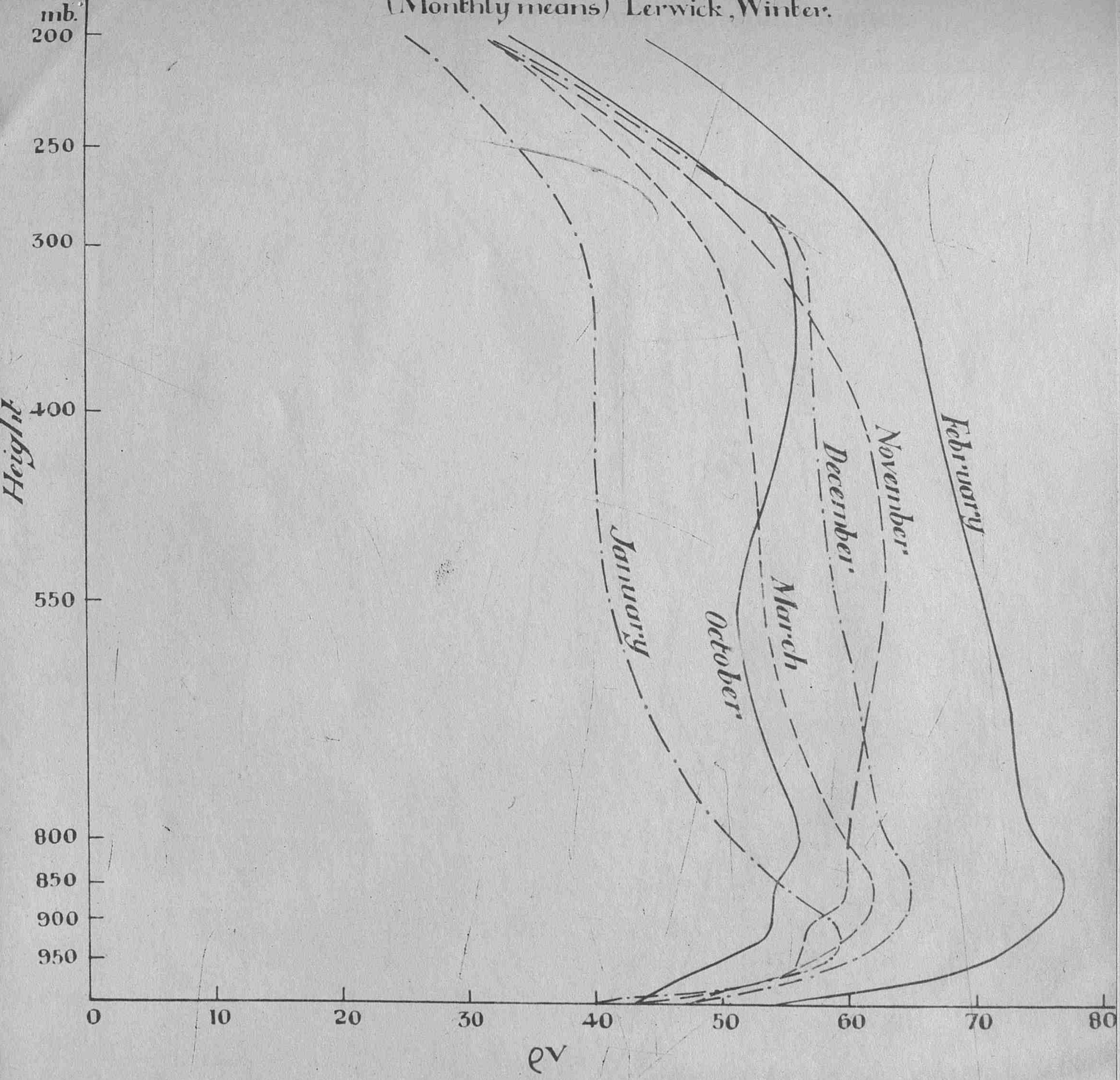


Fig.3a. Variation of Air Mass Transport with Height (Monthly means) Larkhill, Summer.

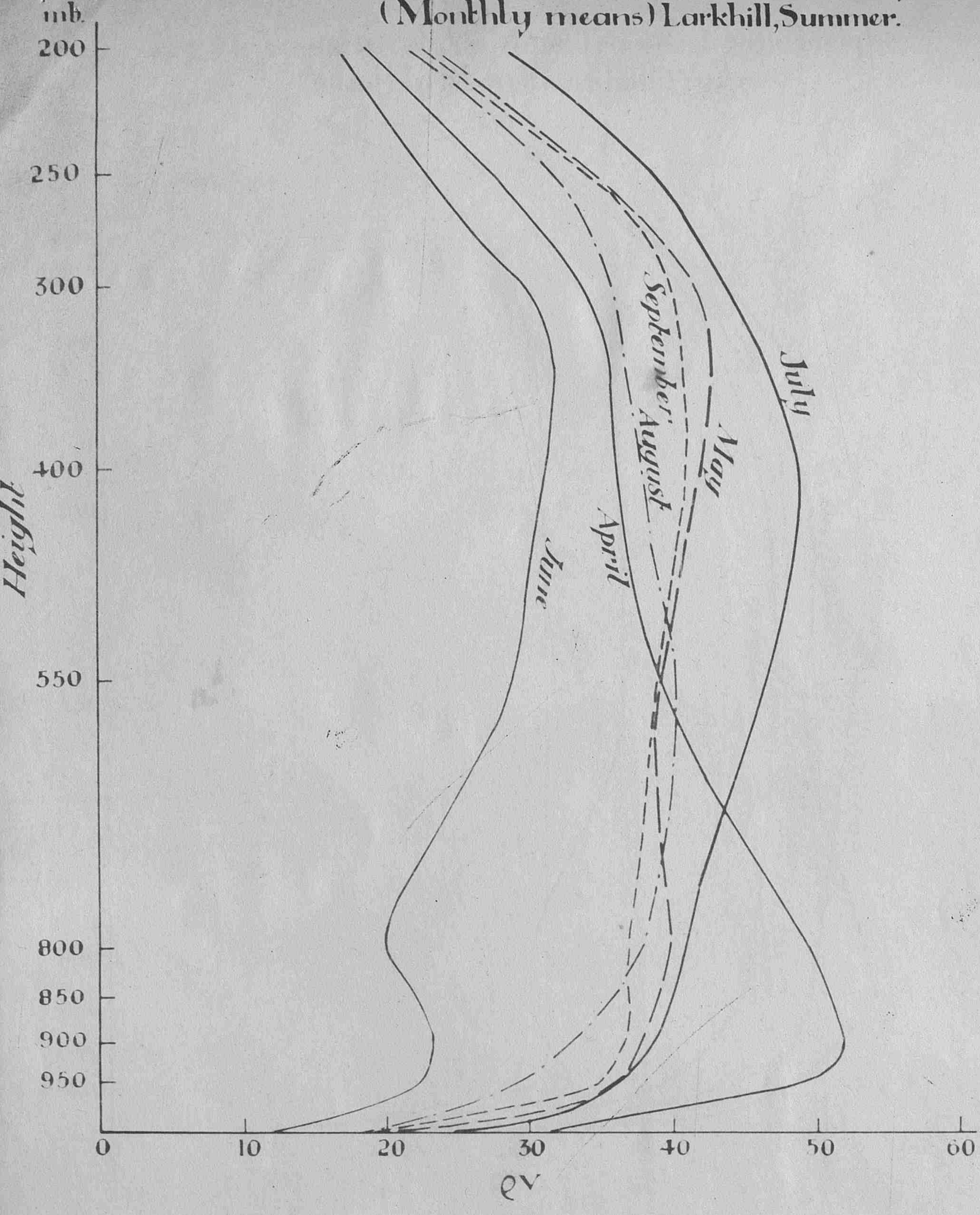


Fig. 3b. Variation of Air Mass Transport with Height (Monthly Means) Larkhill, Winter.

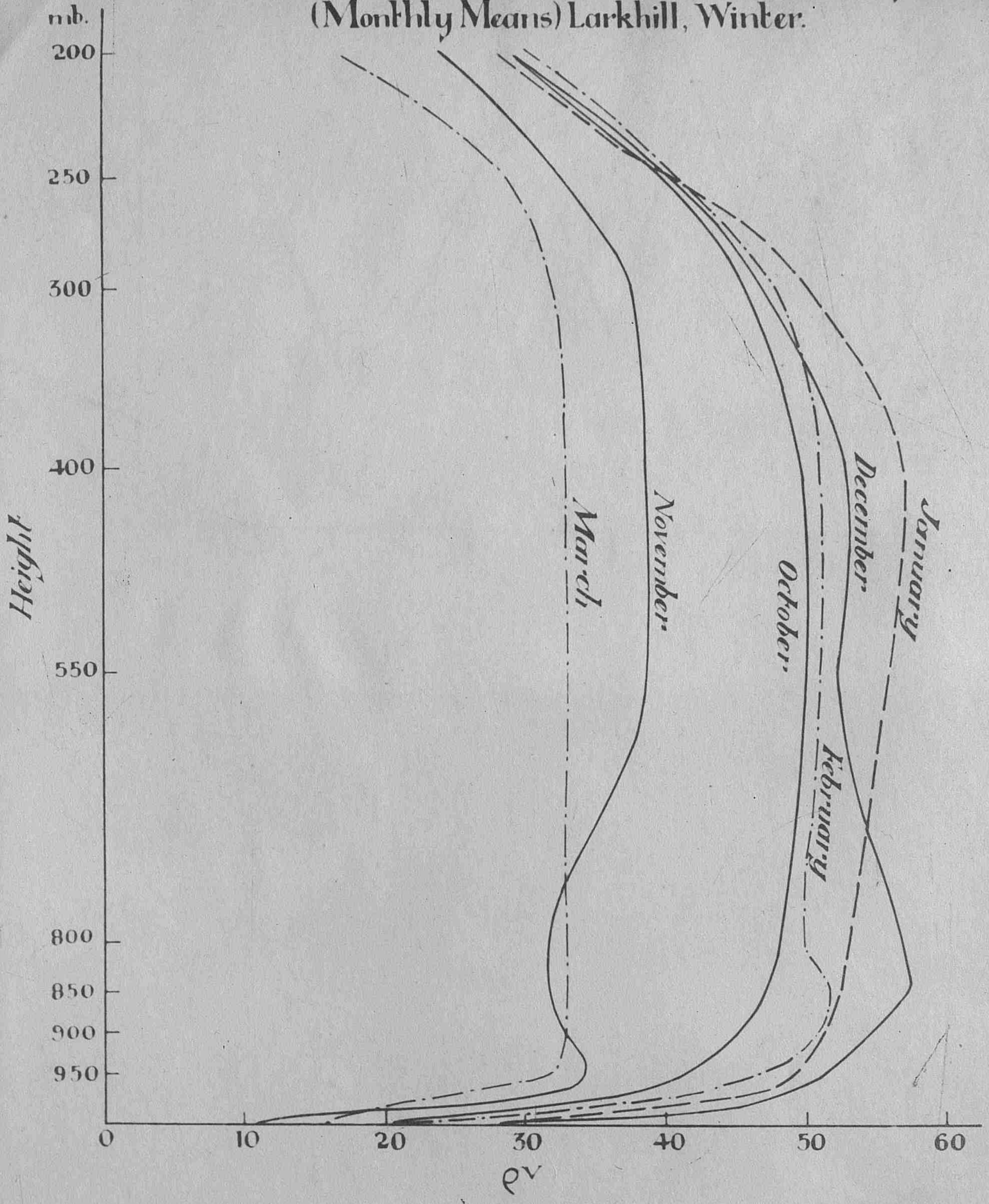
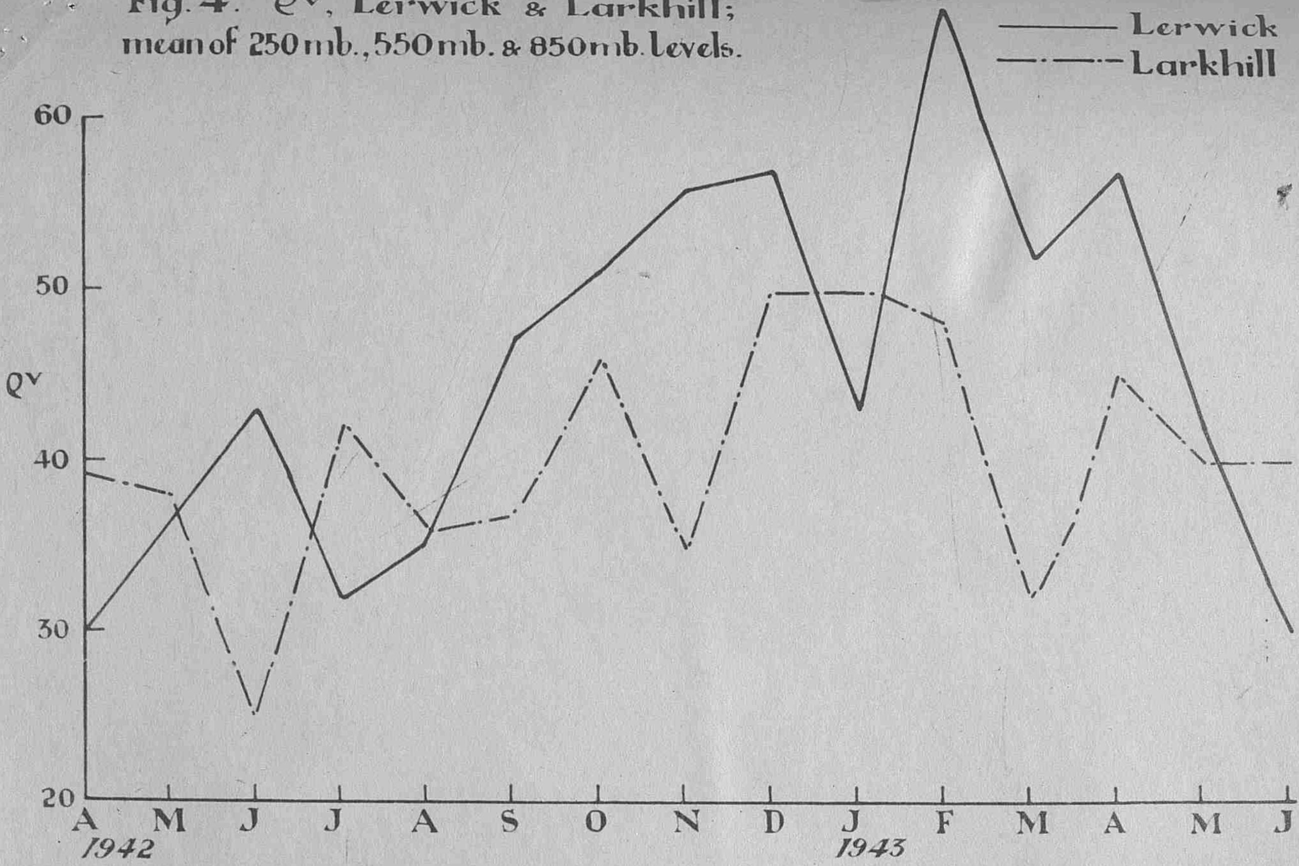


Fig. 4. ρ^{v}, Lerwick & Larkhill; mean of 250 mb., 550 mb. & 850 mb. levels.



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AIR MINISTRY 14th February,1944

METEOROLOGICAL RESEARCH COMMITTEE.

MRP 305 12 Oct 46

8-11-52 ~~E E H~~
Date Initials

Meteorological Information bearing on Aircraft De-icing.

by A.H.R.Goldie.

Introduction. In connection with questions raised by the Meteorological Research Committee, I visited Farnborough on 21st January 1944, and had a discussion with Messrs. Graham, Old, Hales and Hellings. In the course of the discussion, Captain Connor, U.S.A.A.F., who is interested specially in propellor de-icing, arrived.

Mr.Graham first summarized the various expedients adopted to deal with icing. Measures for the reduction of adhesion (Kilfrost paste and slinger ring system) are of some use on propellors, but of much less effect on wings because of the keying action of the ice formed on leading edges. The Goodrich and Dunlop systems were developed in an attempt to cope with this difficulty, but the present feeling is that the thermal method has the best chance of ultimate success. It can also be applied without affecting the aerodynamic properties of the wings.

It is estimated that heat equivalent to from .55 to .70 H.P. per square foot of surface would be needed and that the amount of wing surface requiring heating is about 25% of the chord. The total amount of exhaust heat is enormous but as the proportion to be applied to the wings rises the weight and complication of the apparatus needed to apply it increase in much greater ratio.

It becomes desirable to get as exact information as may be obtained about the maximum possible rate of ice accretion. Since the catch of water drops on the wings depends upon the total free water content of the air at the icing levels and on the size of the water drops, information under these two headings is also desirable; this problem is one which the Meteorological Office already has under investigation.

The heating requirements quoted above are based on an air temperature of -20°C with 1 g/m^3 of freewater, an aircraft speed of 300 f/s and waterdrops mainly of 25μ diameter.

Having regard to the icing casualty rate actually experienced on military aircraft it has been estimated that 400 lb. is the maximum weight of equipment which it would be advisable to add to the Lancaster even if such equipment gave perfect de-icing of wings.

In the case of civil aircraft other considerations enter into the problem and a greater weight of equipment might be permissible in certain circumstances. In this connection it would be valuable to have information as to the extent of the icing danger on various routes, particularly on the North Atlantic.

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Icing of engines, (carburettors, etc.,) may be said to have been overcome and all new engines should be free from this trouble. The same applies to wind screens.

Heating by air or exhaust cannot yet be used for propellor de-icing.

In connection with the matter of adhesion Captain Connor reported that in the case of surfaces treated with a de-icing lacquer it has been found in U.S.A. that whilst the adhesion is low on first formation of ice, it doubles in about 15 minutes if the ice remains in position. He wondered whether the adhesion factor depended in any way on the sublimation nuclei available. Apart from that it seems possible that the increasing adhesion might be due to gradual change from a condition of mixed ice and water towards one of pure ice.

It looks as if assistance in regard to some of the questions now arising could be obtained from observations already made either on aircraft or at mountain observatories. It may therefore be useful to summarize the quantitative information available, more especially since the meteorological notes published so far have been concerned rather with the avoidance of icing conditions than with quantitative data regarding ice accretion, water content, nuclei etc.

1.Amount of accretion on aircraft. W.H.Bigg(1) who examined 217 reports of icing on aircraft suggested that icing might be regarded as slight, moderate or heavy according as the rate of formation was $<$ 0.5in./hr., 0.5-1.5 in./hr., or $>$ 1.5 in./hr. On this scale about 20% of all the reports examined were "Heavy" deposits. Expressed in inches per minute the highest rates reported from aircraft at this
·17/ time(1937) were .25 in./min. in St.Cu cloud, .2 in./min. in ·13/
St.cloud and .1 in./min. in A Cu cloud, all of clear ice. Naturally these relate to experiences lasting only a few minutes since aircraft could not remain long in such conditions.

The rates of accretion at mountain observatories where the facilities are better suited for continuous observation and measurement are dealt with in paragraph 6 below.

2.Temperature Range of Icing. All data examined, both from aircraft and from mountain observations, show the great majority of the cases of icing as occurring within the temperature range of 0°to 32°F (-18°C to 0°C) and mostly in the higher part of this range.

3.Types of Icing. In this country it has been customary to recognize four main types of ice accretion (2), of which the

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first two are opaque and the other two are translucent or clear. These in ascending order of weight are:-

(a) A white semi-crystalline coating, quite light, in the formation of which sublimation plays the chief part; the ground level equivalent is hoarfrost (gelée Blanche, rauhreif).

(b) A white opaque (and not heavy) deposit formed by the freezing of small supercooled waterdrops, the equivalent of rime (givre, rauhfrost)

(c) A transparent or translucent coating which may be heavy, with a glassy surface appearance. It forms within dense cloud consisting of large cloud droplets, generally near the freezing point.
Equal to something between rime and glazed frost.

(d) A heavy coating of clear ice formed all over the aircraft when rain falls on it; equivalent to glazed frost (verglas, glatteis or rauheis).

4. Relation of type of ice to temperature. Bigg found that on aircraft both the opaque and the clear forms give a maximum frequency in the range of temperature 23° to 27°F. but more clear ice cases than opaque occur above 27°F and more opaque than clear ice below 23°F.

Somewhat similar results come from observatories. In the Ben Nevis observations(3) fully 90% of the cases of glazed frost occurred between 28° and 31.9°F. The frequency of saturated or even supersaturated atmosphere was noticeable with clear ice.

In Hilding Köhler's(4) observations in Swedish Lapland the deposits were obtained on spheres on the Partetjakko which, like Ben Nevis, is enveloped in thick cloud during a great part of the year. Köhler's illustrations of the three types which he recognizes are reproduced in Figs. 1 to 3.

The temperatures of occurrence of these types were as follows:-

Rauhreif or hoarfrost. From +14.5°F down to -14°F and chiefly between +10°F and 0°F. (-12°C to -18°C)

Rauhfrost or rime. Over the widest range of temperatures from freezing point downwards at least to -18°F, but more common in the higher range from about +27°F to +10°F (-3°C to -12°C).

Rauheis or glazed frost. In the higher temperatures from +32°F down to +15°F (0°C to -10°C) with an isolated case at -10°F (-23°C).

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B.S.Garonob(5), discussing glazed frost and rime as observed at Odessa found that the glazed ice formed at temperatures between 0° and -2.2°C and the rime at temperatures from -1.2°C to -22.3°C.

5. Relation to humidity and nuclei of sublimation. As to humidity, Köhling(4) found that in all types of icing the vapour pressure was at least equal to that of saturation over ice and in most cases lay between the saturation values over ice and over water surface respectively; in a few cases with glazed frost formation there was supersaturation even with respect to a water surface.

At this point it may be useful to call attention to the difference between the saturations with respect to water and ice surfaces respectively. The difference is shown in Fig.4. It will be seen that it reaches a maximum (of about 0.22 g/m^3) at a temperature of about -13°C(+9°F) i.e. air at this temperature saturated with respect to water - in the absence presumably of suitable sublimation nuclei - could, when brought into contact with ice or probably with any solid object, release 0.22 g. of water per cubic metre and still remain saturated with respect to the ice surface. It may be a matter of accident, but in 8 out of 10 flights made by Mazur (and referred to in para.8 below) the free water contents were .20g/m^3 or less.

If 0.22g/m^3 of free water became available in the air and if it were all caught by a body moving at 200 m.p.h. the maximum ice accretion (measured horizontally) would be approximately 3 inches per hour.

The existence of supersaturation with respect to ice implies absence of nuclei suitable for sublimation. Such nuclei are thought to be of continental origin, probably of quartz , and some evidence recently obtained suggests that in the British Isles supersaturation is in fact found more often in winds which are of northerly or westerly origin (i.e. Atlantic or Arctic origin) than in winds of southerly or easterly (i.e. continental) origin.

Curiously enough much more definite evidence on this point is obtainable from the Ben Nevis work of over 50 years ago.

At Ben Nevis glazed ice occurred on 198 days in the six years 1885-90, with a mean duration of 4½ hours. During the latter half of 1890 an Aitkens dust counter was in use at the Observatory and it was found that the number of dust particles was very small when glazed frost occurred. Further, an analysis made by Mossman(6) of the synoptic weather charts of these 198 days showed that the distribution of pressure was substantially the same in all these cases; there was a depression with its centre off the N.W. coast of Norway and an anticyclone with centre usually over the British Isles or in rare cases in Northern France or further south. In effect Ben Nevis lay in the kind of air current which would be expected to be deficient in sublimation nuclei. Of the 198 cases only 3 occurred in a type of pressure distribution different from the above.

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No case occurred in which pressure was higher to north than to south of Ben Nevis, i.e. no case in "continental air". The precipitation at Ben Nevis at times of glazed frost averaged from .02 to .03 inches per hour and was almost wholly confined to the higher regions.

6. Amount of Accretion as measured at Observatories.

At Ben Nevis(3) the largest growth of ice crystals (on instrument stands etc.) reported was 100 inches in 59 hours or an average of 1.7 inches per hour, with a wind of 80-100 m.p.h. Other observations showed a rough proportionality to wind speed and if this applies up to aircraft speeds (say 200 m.p.h) the rate of growth in this outstanding case would have corresponded to some of the highest experienced by aircraft. At one time Omond attempted to measure the rate of sublimation on a cylinder 3.6 inches long by 2.25 inches diameter and found about .125 inches per hour as the water equivalent of crystals of which he estimated the density as one tenth. This would make the thickness of opaque crystals about 1¼ inches per hour. This was in a wind of Beaufort force 6-8.

Kohler(4) made measurements of the total ice collected in periods of two or three hundred hours on a sphere of 20 cm. diameter. The figures given are the volume of water yielded by the deposit when melted. If the volume collected is divided by the cross section of the object exposed to the wind, the accretions so derived in the different months range up to about .8 mm. per hour in a wind averaging 10 miles per hour. This (raised to an airspeed of 200 m.p.h) would correspond to uniform clear ice accretion of 16 mm. or 0.64 in. per hour.

In 1935 at Mount Washington Observatory S. Pagliuca(7) made a series of measurements of rime accretion on a model aerofoil exposed as a rule for several hours at a time in winds ranging from about 50 to nearly 80 m.p.h. He measured the thickness of deposit and also the density of the ice. The densities varied from .27 to .96. In a typical case, with a density of .90 the deposit at 75 to 80 m.p.h. was 106 mm. per hour corresponding to about 1.10 in. per hour, at 200 m.p.h.

If the figure derived from Köhler's water equivalent is raised to correspond to ice of density .90, it would become .71 in. per hour. Since however this is the average over a surface on which the actual ice accretions were by no means uniform, it is, as would be expected, still somewhat lower than the values derived by Pagliuca's method. It will therefore be noted that the values derived both from Köhler's and from Pagliuca's mountain observations correspond to about the middle of the range termed "moderate icing" by Bigg in the analysis, referred to in para.1, of the aircraft observations.

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7. Icing in relation to size of obstacle and size of water drops.

The rates of accretion depend not only on wind speed, but also on form and dimensions of the obstacle, as well as on the size of the water drops.

For a given size of obstacle (Köhler's observations were made with spheres) the shape of the accretion depends on the type of icing and that in turn on whether drops are present and on their size, other conditions being equal. The larger the obstacle the greater is the thickness of the stream of diverted air through which a drop of water has to pass to impinge on the obstacle; and the smaller the drop the more likely is it to be diverted by the airflow. This has been treated mathematically by various writers(8).

As Figs. 1 to 3 show, the hoarfrost (Rauhreif) is essentially a sublimation. The rime (rauhfrost) takes the form of blunted cones pointing into wind, shortest on the centre of the spherical surface which faces the wind and longest round the periphery, so that the aspect presented to the wind is concave. This arrangement results from the fact that the small water drops which produce rime tend mostly to be deflected and to follow the stream lines of the air as it passes round the sphere, so that the longest cones are built up around the periphery. (A corresponding formation is sometimes observed on aircraft)

The glazed frost (rauheis), being formed by larger drops whose momentum carries them straight on, builds up in a shape which is more or less parallel with the windward surface of the sphere and is thus convex towards the wind.

Larger obstacles catch relatively little deposit on the lee side as compared with smaller obstacles. Köhler found that a sphere of 20 cm. diameter acquired a total weight of deposit of only twice as much as a sphere 10cm. in diameter; Albrecht(9) at the Sonnblick Observatory in Austria carried the investigation further using obstacles of relatively small size. He exposed lengths of wire of from 0.5 to 6 mm. diameter. The ice accretions of a simultaneous 30 minute exposure of these wires showed relatively small accretions on the thicker cross sections. (Noth and Polte(8) state that monoplanes with a thick wing profile accumulate ice more slowly than biplanes with thin wing sections.)

8. Water drop sizes and free water content of clouds.

The drop sizes on the occasions of the above measurements by Albrecht were estimated as being 26μ diameter and the water-content as 0.91 g./m^3.

In another investigation Köhler(10) found that water drop sizes measured from optical data agreed with the sizes as measured microscopically of the drops which resulted in icing. He found that in clouds and mountain mists the drop sizes sometimes changed considerably within a very short time, the radii in general remaining within the range 3 - 25μ, though a few measurements gave larger or smaller sizes. On other occasions the measurements for several hours gave, within the limits of observational error, the same size. But it also

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happened that the drop size, after remaining constant for a time, then suddenly changed. The new size would be constant for a time and then again might change, sometimes back to the original size. The masses of the measured drops always fell into the following series:-

2 : 3 : 4 : 5 : 6 : 7 : 8 : 10 : 12: 14 : 16 : 20 .....

with the first number always indicating the same mass, independent of temperature or place of observation. Niederdorfer, within the errors of measurement, found the same mass series. The most frequent size of droplet in clouds causing icing had a diameter of 17.64μ.(11)

In an investigation arranged within the past year by the Meteorological Office, Mazur(12) by means of an "aircraft impactor" has taken samples of water drops in clouds and estimated their sizes and frequency distribution. The results, which are given in detail in M.R.P.109, may be summarised as follows: The curves which represent the drop size distribution are very similar to a Gaussian or normal distribution curve. Each has a single and very strongly marked maximum, a definite and rather sharp lower limit to drop sizes and an almost indefinite tapering away towards the higher sizes. The diameter of the smallest was less than 1μ and of the largest about 100μ . The region of maximum frequency in the majority of some 36 curves lay between 10 and 20μ which seems to be in agreement with results from mountain observatories. The free water content of the clouds explored by Mazur in ten flights lay in 8 cases between 0.12 and 0.20g/m^3. In the remaining two cases (St.cu. and Cu. respectively) the free water contents were 0.38 and 0.41g/m^3.

Explanation of Figures 1-3.

1. (1) and (2) Rauhreif
   (3) and (4) Rauhreif changing over to Rauhfrost.

2. (1) Rauhfrost, beginning of formation.
   (2) Rauhfrost.
   (3) Rauhfrost viewed from windward.
   (4) Rauhfrost - the same viewed from one side.

3. (1) Beginning of rauhfrost formation.
   (2) Rauhfrost and transitional form to rauheis.
   (3) Rauheis viewed from windward.
   (4) Rauheis - the same viewed from one side.

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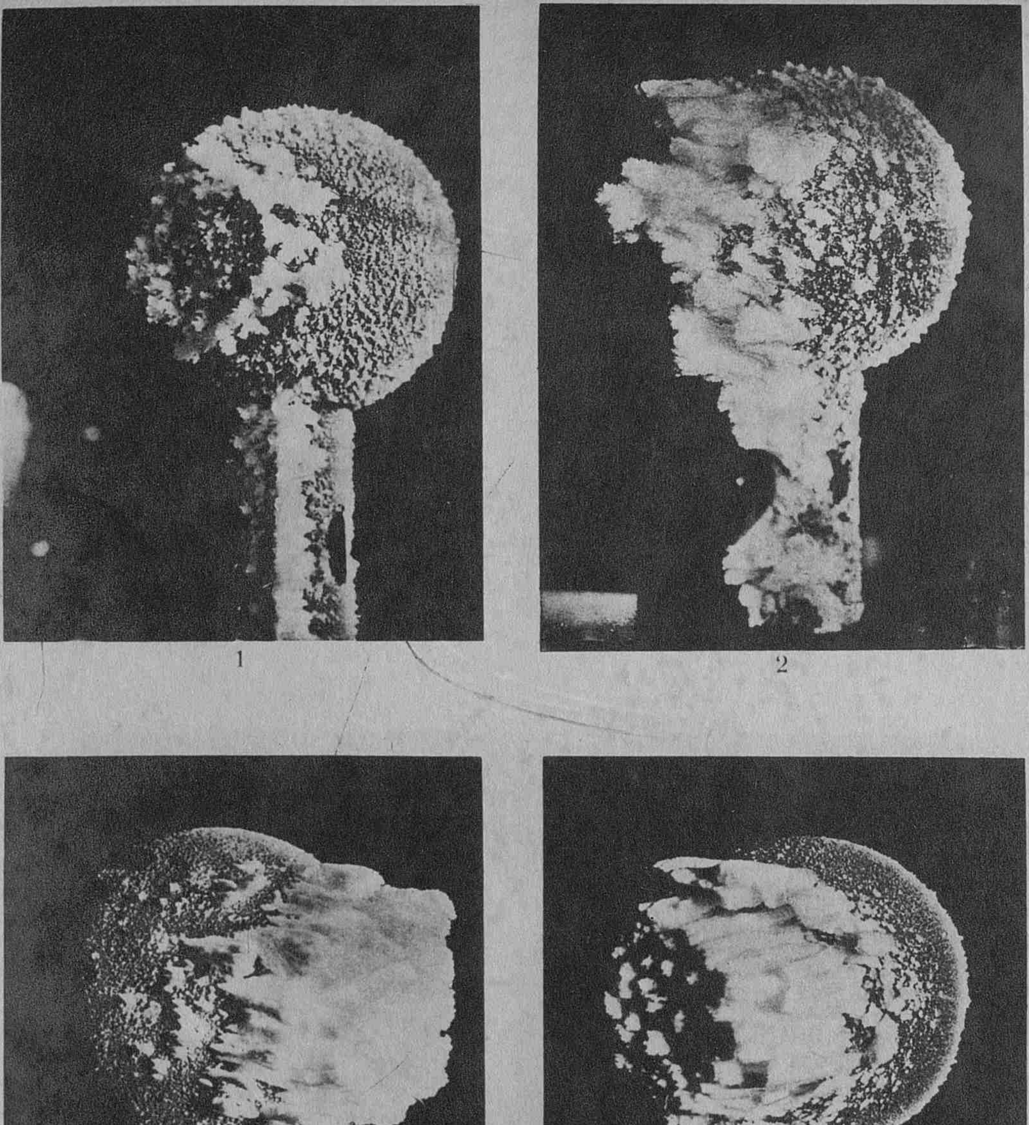

1 2

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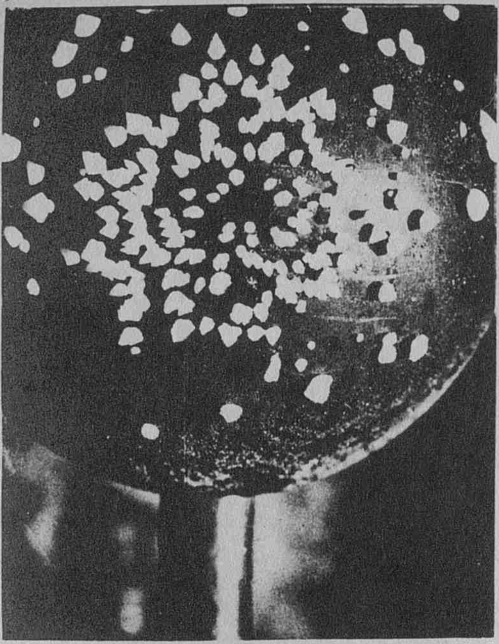
H. Köhler phot.

Figure 1.



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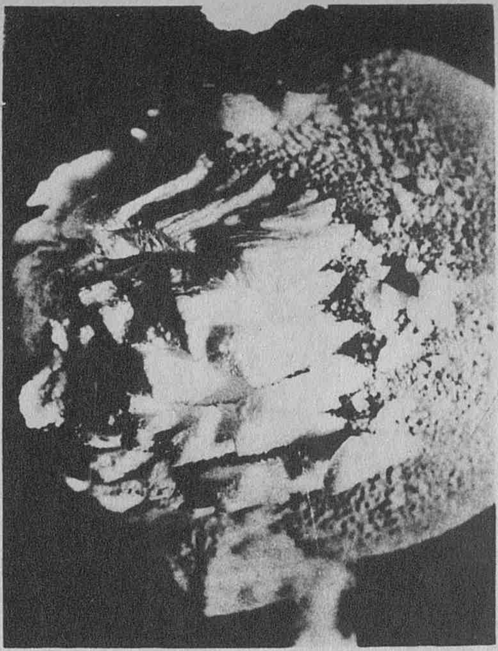

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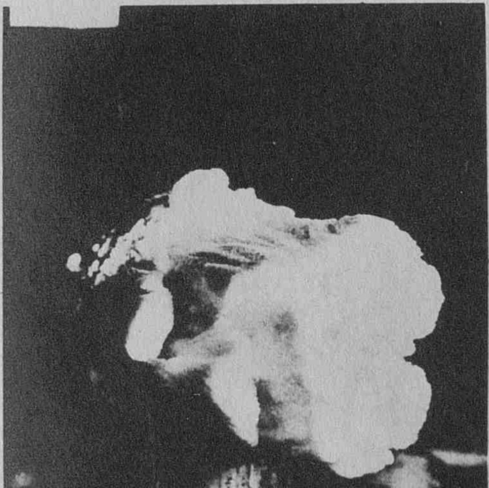

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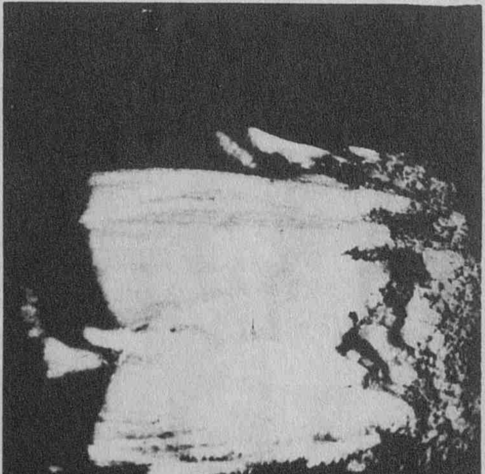

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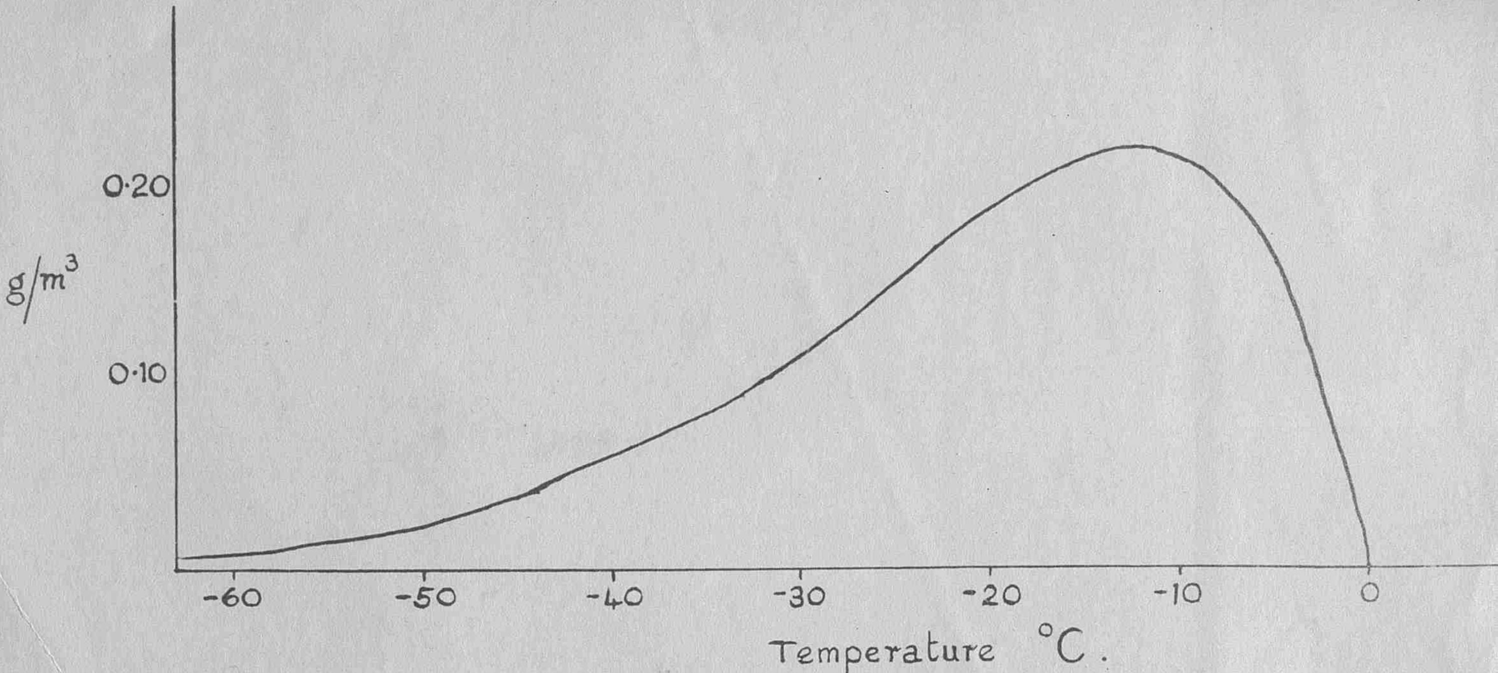
H. Köhler phot.

Figure 2.

Figure 4 Difference in water content between saturation with respect to water & ice surfaces.



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MRP. 169
14th February
1944.

Work carried out on the Fortress Aircraft Allotted to H.A.F. for Meteorological Duties

by A.W. Brewer.

1. Introduction.

1.1 In June, 1943, the Fortress Aircraft F.K.192 was allotted to H.A.F. for Meteorological Duties in continuance of the work on the "Dew Point" Hygrometer designed by Dr. Dobson F.R.S. This work was begun on the Boston aircraft formerly allotted and was described in a report upon the hygrometer (M.R.P. 126). The earlier work showed that with care the hygrometer was capable of giving consistent and reasonable readings down to frost points of about -50°C, and should work, though with greater difficulties, down to the lowest temperatures encountered in the atmosphere, about -80°C. Further development was contemplated on the hygrometer to improve its performance and facilitate its use.

1.2 The Fortress aircraft, capable of climbing about 10,000 feet higher than the Boston, that is to 37,000 ft., was therefore allocated with three specific problems in view.

1. Continuation of the development of the hygrometer for the very high altitudes.
2. Investigation of Condensation trails.
3. Investigation of the water content of the upper troposphere and lower stratosphere.

In view of the fact that this was the first time that a hygrometer, capable of use at the low temperatures involved, was available, the field of investigation of items 2 and 3 was new ground and striking results could be expected.

? 1.3 Since the aircraft arrived at Boscombe Down, about 90 hours flying have been done, of which about 70 hours have been on work directly connected with hygrometry. The flying time has been restricted partly by unserviceability and partly by the fact that in large aircraft of this kind climbs above 10/10 cloud have not so far been permitted. Most of the unserviceability has been due to two minor accidents, one when a wheel broke up apparently spontaneously as the aircraft stood on the ground and the other when an aircraft taking off just hit the top of the tail fin and caused minor damage which took some time to repair. Since November 1943, we have also had a lot of trouble with the cabin heating. It is not practicable to climb this aircraft without some heat.

1.4 At first considerable difficulty was experienced in finding a crew all of whom could withstand the physical strain of the climbs. The crew must consist of a Pilot, Flight Engineer, a rear Observer to observe trails and who is also, preferably, a Wireless Operator, and a Meteorological Observer. The choice of personnel capable of carrying out the technical duties and withstanding the physical strain of the climb is very limited. The usual procedure is to climb to ceiling, about 36,700 feet, without interruption. This takes about one hour. Observations are then made on the descent every 20 mb. between 220 mb. and 300 mb., and then at the following pressure levels: 325, 350, 400, 500, 600, 700, 800 and 900 mbs., the whole flight taking about 2½ hours, most of which is above 30,000 feet (300 mbs.) Experience has shown that it is unwise to make the observations on the climb. This is partly because the levels involve extensive adjustment of engine controls and also because by the time a high altitude is reached, the crew have become too cold and exhausted to persist in the climb to the ceiling. It may be mentioned that on account of freezing of the elevator trimmers the physical effort called for from the pilot is considerable.

1.5 As received at Boscombe Down, the aircraft had a ceiling of about 34,000 feet. This has been improved to a value well over 37,000 feet, by the removal of all guns, armour plate, turrets and the reduction of the normal fuel carried from 1,000 gallons to 700 gallons.

2. Expenditure of Flying Time.

2.1 Of the 70 hours flying time an appreciable amount has been consumed in abandoned climbs from which no useful results were obtained. Flights have been abandoned for the following reasons:-

(a) Personal troubles.
  1. Bends (Mr. Brewer due to unserviceable oxygen mask)
  2. Stomach Pains (Pilot)

(b) Aircraft Troubles.
  3. Intercom. failure (several flights)
  4. Engine failure (surgeing of No.2 Engine)
  5. Cabin heating failure (several flights)
  6. Engine Instrument failure (boost gauge)
  7. Hygrometer failure (Blocking of the cooling jet)

Also three climbs have been made from which no useful results have been obtained, testing new and unsuccessful hygrometers. Two of these were made with the original "perspex ellipse illumination" hygrometer, which had unsatifactory jet action. The third was with a second "perspex ellipse" hygrometer. On this flight it was not realised that the perspex was giving up water, causing hoar frost to be deposited densely on the thimble outside the jet. Subsequent flights showed that it was quite practicable to use this hygrometer if the deposit outside the jet were ignored.

2.2 Of the useful flying time about 2½ hours have been spent checking the speed corrections of the thermometer and general accuracy of the thermometry, and the rest to climbs, mostly made in accordance with the procedure of 1.4 above.

3. Hygrometer Installation.

3.1 The installation, at the time of writing, in the aircraft is shown in the photograph, Fig.I. This is a temporary installation which permits the comparison of two very differently constructed hygrometers.

3.2 Both instruments are built into the navigator's table and outside air is supplied through the half inch copper tube which stretches from the instruments to the centre of the window. When only the instrument is installed (the normal arrangement) part of the air passing down this copper tube is allowed to escape into the cabin in order to ensure that

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this pipe is well ventilated and so to reduce trouble from adsorbed water upon its surface. The tube has also been cleaned in chromic acid to reduce this adsorption. The thermos flasks which contain the coolant are supported below the table by strong rubber bands.

3.3 The two instruments are very different in detail. In the right hand instrument the thimble is of aluminium, anodised and dyed black, and illumination is by means of a perspex disc. The thermometry is carried out by means of a nickel resistance thermometer wound in the grooves of a fine thread cut on the skirt of the thimble and insulated by means of the anodising. The temperature is read off by means of the balanced bridge which may be seen to the left of the hygrometers. By means of a simple change over switch this same bridge can be used to measure the outside air temperature. The outside air temperature element is a platinum resistance thermometer. The bridge is calibrated to read directly the temperature of this air temperature element and corrections have to be applied to the thimble temperature readings, partly on account of the different metal, and partly because the thimble bridge is adjusted to cover a larger, and lower, temperature range. The thimble temperature of course enables the dew point to be obtained directly without speed or other corrections. This arrangement also has the advantage of eliminating the rather delicate microammeter necessary for use with thermo-couples, and it is very much faster in response than the thermo-couples. This last point is very important.

3.4 The left hand instrument is the hygrometer used in the earlier test work on the Boston aircraft and described in the report on that work. The deposition surface is gold, the illumination is by mirrors bent approximately into elliptical shape, and the thermometry is by thermo-couples.

3.5 It may be mentioned that the first instrument is better as regards the illumination and contrast of the dew and the thermometry, while the second is better as regards the action of the jet and absence of trouble due to absorption of water by the perspex.

3.6 The external air scoop and the external pitot head for the thermo-couples are seen in the photograph, fig.2. When desired the scoop and pitot head can be protected by means of the sliding shield which in the photograph is retracted to expose them.

4. Results of Speed correction trials.

4.1 To examine the relations between the various thermometers, their speed corrections and agreement with each other, a speed trial run was carried out under 10/10 turbulent cloud. The thermo-couple hygrometer was inverted, the thimble filled with a light oil, and a standard thermometer inserted in the oil. With the thimble well lagged, the temperature was steady enough to be confident of the thermometer indicating the thimble temp to about 1/10°C. Then by means of the thermo-couples connected differentially to the external pitot head and the microammeter it was possible to use the system as an air temperature thermometer with a known speed correction of 100%. The other thermometers on the aircraft were two strut psychrometers, one Mk.II and one Mk.V, and a balanced bridge resistance thermometer with the element used in the radiation shield which is now standard for Prata aircraft. The element for this thermometer was first mounted on the belly of the aircraft about 8 feet from the nose, but trouble was experienced with it reading up to about 1°C. different from the strut psychrometers and the element was moved forward right up to the nose of the aircraft. The differences then disappeared.

4.2 The results of one speed run are shown graphically in Fig. 3, in which the differences between the different thermometers are shown plotted against $\frac{(V\ mph)^2}{100}$. The agreement between the various thermometers is satisfactory. The results show the speed corrections for the various thermometers to be as shown below, for which the value of α is given for each thermometer to fit the formula $T = T_0 - \alpha \frac{(Vmph)^2}{100}$ °C.

Thermometer	α
Forward facing pitot head	1.00 standard
Mk.II Psychrometer	.80
Mk.V "	.70

4.3 During part of the period under review the balanced bridge thermometer was sent away to be copied by a manufacturer, so that for a few climbs there was no thermometer available reading below -40°C. For these climbs the air temperatures of the Larkhill radio sonde have been used, for consistency for the whole climb. The hygrometer indicated the dewpoint depression so that any error in the temperature is not reflected in the humidity.

5. Results of Ascents - Aerodynamic Condensation Trails.

5.1 The Fortress aircraft makes trails in three ways, the familiar exhaust trail, which will be dealt with in detail later, a condensation from the propellor tips and also a condensation on the upper surfaces of the wings. These last two are often visible as the aircraft passes through moist air and on damp mornings the propellor tip condensation can be seen when the aircraft is run up. They are normally highly ephemeral but in supersaturated air the condensation on the top of the main planes can form a persistent trail. This type of trail was experienced in the Boston where it seemed to come from the propellor tips but in the Fortress the wings seem to make the main contribution. In the Boston condensation on the upper surfaces of the wings was never noticed.

5.2 These aerodynamic trails have been made twice in the Fortress. Once was before instruments were fitted and the second occasion was on 5th January 1944. As on the occasions of the formation of aerodynamic trails in the Boston aircraft, the trails were made in the neighbourhood of the broken cloud sheets of weak or dying fronts.

5.3 On 5th January at 12,000 feet, at a temperature of -13°C (8°F), the trails were made continuously just beneath a very well broken layer of Alto-cumulus. The aircraft circled for some time while humidity observations were made though, at that level and temperature, the cooling is rather violent. The hygrometer indicated a dew point depression of 0.5°F, which is reasonable in view of the fact that we were immediately below Altocumulus cloud which corresponds to a supersaturation over ice of about 10%. The trails formed a dense pattern in the sky and about ½ hour later the Altocumlus patches had dispersed and the trails were dispersing by falling out as snow. They vanished entirely by this mechanism in about ¾ hour.

6. Results of Ascents, Persistent Exhaust Trails.

6.1 Exhaust trails are to be divided into the two main types of persistent and non persistent trails. The truly persistent exhaust trails which persist for a long time and become Cirrus cloud have been made on two ascents in the Fortress and on these occasions supersaturation with respect to ice has been indicated by the hygrometer. Supersaturation with respect to ice has been observed on one occasion without any trail being formed. Details are tabulated below:-

Date	Ht.	Pressure	Temp. °C	Frost point °C	Supersaturation %	"Mintra" °C
		Persistent Trails formed				
7.9.43	32,000	274 mbs.	-49	-48	8%	-39
"	31,000	286 "	-45	-44	10%	-39
"	28,500	322 "	-39	-37	20%	-37
24.8.43	31,000	287 "	-49	-48	10%	-39
		No Trail Formed				
24.8.43	28,800	332 mbs.	-40	-39.5	5%	-37

This lists all occasions in which persistent trails have been formed and supersaturation observed.

6.2 The last observation in the table at 27,800 feet on 24.8.43 is interesting as no trail was formed either by the Fortress or by an accompanying Spitfire. The supersaturation is slight and may not bevreal and evidence will be given later to suggest that the Mintra is perhaps about 4°C too warm.

6.3 The climb on 24.8.43 was the only one on which a co-operation of a Spitfire has been obtained. The slow rate of climb of the Fortress and the short endurance of the Spitfire are the main difficulty. A "rendez-vous" was arranged with a Spitfire at 30,000 feet on one occasion but we failed to meet.

7. Non Persistent Exhaust Trails.

7.1 Goldie has shown that below a certain temperature a trail is to be expected according to the relative humidity. The relation between this temperature and height have been given in M.R.P.116. This temperature is commonly called the "Mintra" temperature. The evidence given above suggests that the persistent trail is formed only under conditions of complete saturation or supersaturation with respect to ice.

7.2 Non persistent trails have been formed on all but one of our ascents. On the one ascent the stratosphere was so warm and dry that none was formed even in the stratosphere.

7.3 There is no exact time of beginning of the formation of trails as when conditions are near the boundary they can be induced by varying the mixture or by opening the throttle which increases the water vapour output while there is a lag in the heat output, so that the trail formation will depend on the past history of the aircraft and its engines. Nevertheless it is possible to show roughly the relation between temperature and dew point depression which separates conditions in which trails will form and will not.

7.4 In the table below are shown for the various ascents (a) the first or last observation in which a trail was formed and (b) the first or last observation of temperature and humidity of occasions when no trail was formed even though the temperature was below the "Mintra" of Goldie. The observations are made at varying heights and to reduce them all to comparable figures, the difference between the temperature and the "Mintra" temperature is also tabulated.

First or last observation with trails forming.

1 Date	2 Height	3 Temp °C	4 Description of trail.	5 Temp.dep. below Mintra. °C	6 Dew point depression °C
26.8.43	29,300	-45	Non persistent	7.3	4.5
6.9.43	30,600	-45	Very short	6.6	6.6
7.9.43	28,500	-38	Persistent	2.2	-1.7
7.9.43	30,000	-43	Short	5	3.4
24.8.43	28,800	-43	Non persistent	5.6	2.2
5.12.43.	28,300	-46	Very short	9	9
22.12.43	26,600	-43	"	6.6	2.2
22.12.43	31,600	-50	Moderate	11.1	16.9
24.12.43	28.300	-48	Short	11.7	7.2
14.1.44	28.300	-49	Very short	12.8	7.2
4.2.44	26,600	-41	Short	4.5	-0.5
4.2.44	28.300	-44	Short	8.3	7.7

First or last observation without trails.

1 Date	2 Height	3 Temp °C	4 Description of trail.	5 Temp.dep. below Mintra. °C	6 Dew point depression °C
22.8.43	30,000	-45		7.5	13.0
24.8.43	28.000	-40		3.3	- 0.5
6.9.43	30,000	-44		5.9	4.4
3.10.43	30,000	-45		7.2	9.0
9.10.43	30,000	-44		7.2	14.0
5.12.43	26,600	-41		5.0	9.0
22.12.43	28.300	-45		8.3	6.6
22.12.43	33,100	-53		9.0	23.0
24.12.43	26,600	-44		7.8	10.0
14.1.43	26,600	-45		9.0	7.2
4.2.44	28.300	-44		8.3	10.0

Observations were originally reduced to the nearest whole degree Fahrenheit and in the conversion in columns 5 and 6, the whole deg.F. have been reduced to the nearest 1/10°C. The probable error of the air temperature measurement is about $\frac{1}{2}$°F and the dew point 1 or 2 degrees F., becoming greater at lower temperatures.

7.5 These observations are all plotted in Fig. 4 in which the temperature depression below the "Mintra" temperature is plotted against the dew point depression. The first or last occasions where a trail was formed are marked X and the first or last occasions where no trail was formed are marked with a ʘ It will be seen that though there is a fair amount of intermingling the dotted line devides the points roughly into occasions of trails and no trails and points which are on the wrong side of the line are never far over it, especially having regard to the difficulty of the humidity observation. There are two main features about this curve, firstly its intercept on the zero dew point depression (100% humidity) line, which is at about 4°C suggesting that the Mintra temperature of Goldie is about

4°C too warm and secondly that it appears to be assymptotic to a line corresponding to a temperature about 13°C below the Mintra temperature. This implies that even in completely dry air a trail is formed if the temperature is about 13°C below the Mintra. It will of course be short. This means that the equivalent cross section area of the trail for a short trail is less than the 30 square metres which Goldie assumed for a Spitfire. For a trail of cross section area of 30 sq.m ., which Goldie assumed, the curve would become assymptotic to a line about 20°C below "Mintra".

8. Results. Water Content of the lower Stratosphere.

8.1 The Stratosphere has been definitely reached and entered on five of our ascents. The main feature of all the ascents is a sudden and very intense fall in absolute (and therefore in relative) humidity as soon as the stratosphere is reached. The only two occasions on which we have got more than 3000 feet in the stratosphere, the air has dried beyond the range of the hygrometer and no deposit could be obtained. See figs. 6 and 7 for T.φ grams of two ascents reaching the stratosphere.

8.2 To illustrate this drying effect the results of all ascents entering the Stratosphere have been plotted in fig.5 on a temperature height diagram but, to reduce all the ascents to a comparative condition, they are all plotted with the tropopause height and temperature shifted to the origin. It will be seen how in every case the dew point is falling off very rapidly within the stratosphere and at a rate without sign of diminution at the top of the ascent. On 22nd December (see fig. 6) at 6000 feet. into the stratosphere no deposit was obtained with a thimble temperature of -80°C, which implies a humidity of 1% or less - drier than anything yet encountered in the troposphere.

8.3 A non-saturated stratosphere has always been expected even on the assumption of a saturated tropopause and was confirmed by the normal reduction of condensation trail length on entering the stratosphere, but nothing of this sort has been visualised before. It cannot be said yet whether it is an essential feature of the stratosphere as these observations are too few and are all restricted to low tropopauses.

8.4 The question of the water vapour content of the stratosphere is very important to theoretical Meteorology, for the problem of the radiation balance of the stratosphere is still unsolved. Simpson assumed a water content of the stratosphere of 0.3 mm of water. This certainly would not be attained under conditions shown by the ascents. With a dew point of -80°C the water content is reduced to only a fraction of the carbon dioxide content at ground level.

9. Results. The water content of the tropopause.

9.1 The results of ascents made in the Boston aircraft and quoted in a previous report showed the extreme variability of the humidity of the upper air, where the humidity could vary between 10 or 20% supersaturation with respect to ice mainly in regions of dying fronts, to a relative humidity of 2% or 3% in strong subsidence regions. The results obtained in the Fortress have continued to show these features and also have shown the value of a complete ascent with the water content shown through the whole atmosphere from the earth's surface into the stratosphere. The value is enhanced by the direct display of the dew point which is a fundamental and often very conservative thermodynamic variable.

/9.2

9.2 Probably the best example which we have yet had is shown in fig. 6 where the ascent of 22nd December 1943 is plotted on a temperature - Entropy diagram. Note the deep layer extending from 700 mbs. (10,000 feet.) to 325 mbs. (28,300 feet) with constant water vapour content of .07/grm/m^3 and a humidity increasing from about 3% at the bottom to about 95% at the top. The implication that subsidence has occurred through this immense depth is very strong. Presumably the process has continued for a fair period of time to permit cooling by radiation of the air, which, at the base, has a potential temperature 19°C lower than at the top, though so much cooling by radiation is difficult to understand.

9.3 Another example worth examination is the ascent on 7th September 1943 and shown as a TØgram in fig. 7. This ascent again shows a dry stratosphere but with a supersaturated layer beneath extending from the tropopause (250 mb.) to about 350 mb., and then there is a dry layer extending from 400 mbs. to 700 mbs. The observations would be quite in accordance with the idea that this layer had descended en bloc through about 200 mbs. The origin of this air judged from the Upper Air Section of the D.W.R. is between 40°N and 50°N., so that it may have originally been air beneath a tropopause at about 200 mbs. which has been pushed down by the introduction of the frontal air above it. The change from supersaturation at 330 mbs. to 4% R.H. at 400 mbs. is probably to be explained in part by the falling out of snow from the upper layer.

9.4 It is unfortunately impossible to treat these very isolated ascents synoptically and any deductions must be regarded as tentative.

10. Conclusion.

10.1 All the above demonstrates the value of the new weapon of a hygrometer which operates at all atmospheric temperatures as a means of attack upon meteorological problems. The "dew" point hygrometer is of particular and unique value in dealing with the low humidities of the subsidence layer in which if a hygrometer of the relative humidity type were employed much would be missed. For instance the significance between say 2% and 5% relative humidity could not be relied upon but the corresponding 8°C difference in the saturation temperature may be very significant and could be reliably measured.

10.2 New problems arise which may be attacked, but large numbers of these ascents, not 10, are required. Such problems are:-

a. Stratosphere.

1. Is extremely low relative humidity usual or is it always present in the stratosphere.

2. How are the low relative humidities which have been observed to be explained.

3. If very dry stratospheres are the rule in these latitudes, are they so in equatorial regions. If so, how do they arise.

4. If the low humidites are due to a photochemical process how is it that very high relative humidities or supersaturation appear stable just below the tropopause.

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5. It is possible that interchange of air between the stratosphere and troposphere is only on a very small scale in spite of the large scale vertical movements which occur in the troposphere and possible in the stratosphere.

6. Data on the water and CO_2 content of the stratosphere are required for the examination of its relation to the radiation balance.

b. In the tropopause there can be no doubt that if such observations were available on a synoptic basis, much progress could be achieved.

10.3 In reviewing what has been done it seems fair to say that the work on the Fortress for all its disappointments as regards serviceability, the restriction on its use above cloud, of the large crew required, and its limited ceiling; has yielded good dividends and promises great things for the future. We look forward to the day when a smaller pressure cabin aircraft can be obtained for the work, such as a Welkin or suitable Mosquito. Complete redesign of the hygrometer would be required but the additional 8000 feet or so in attainable ceiling would permit reaching the stratosphere on almost every occasion at any latitude.

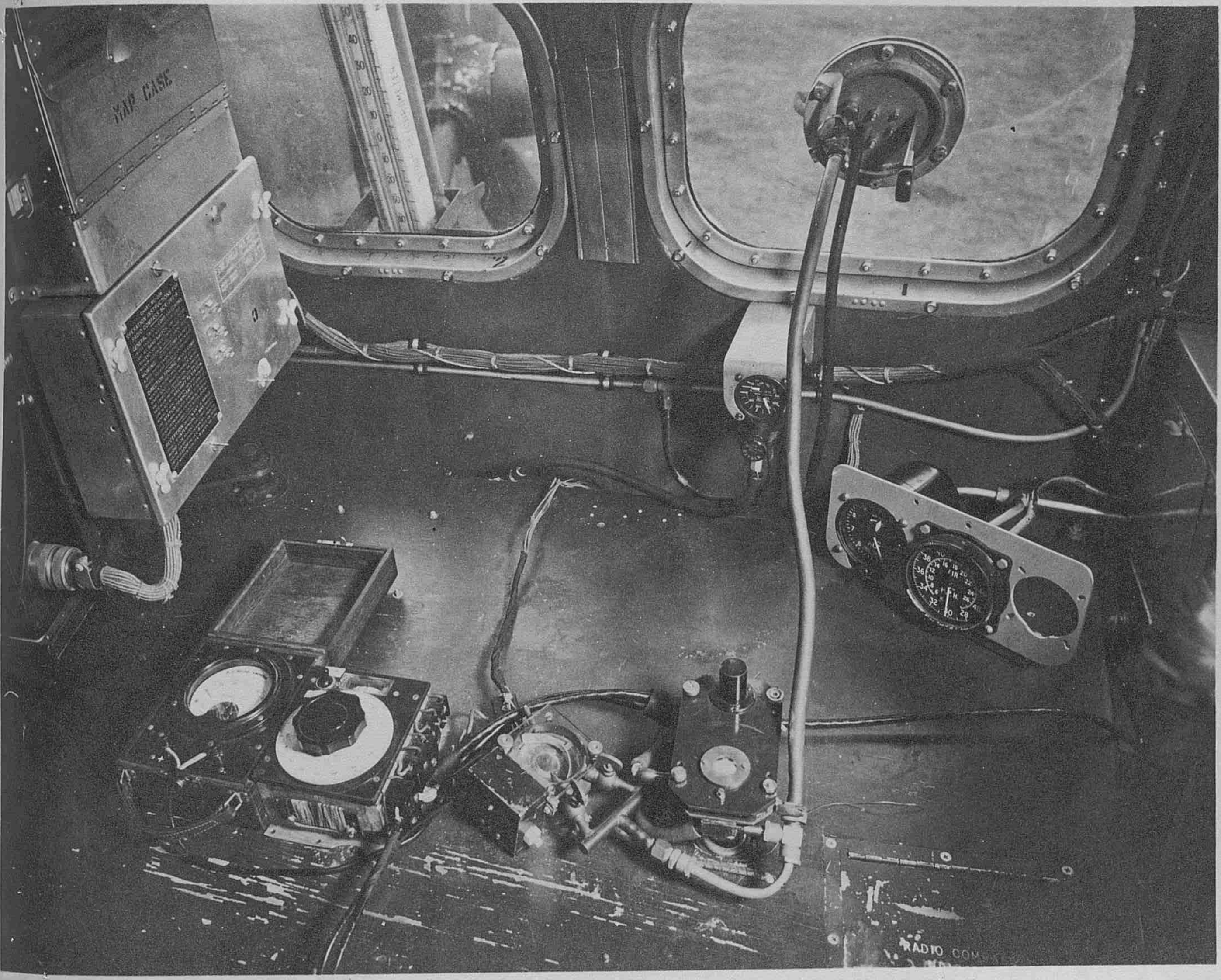

MAP CASE
RADIO COM

Fig. 1.

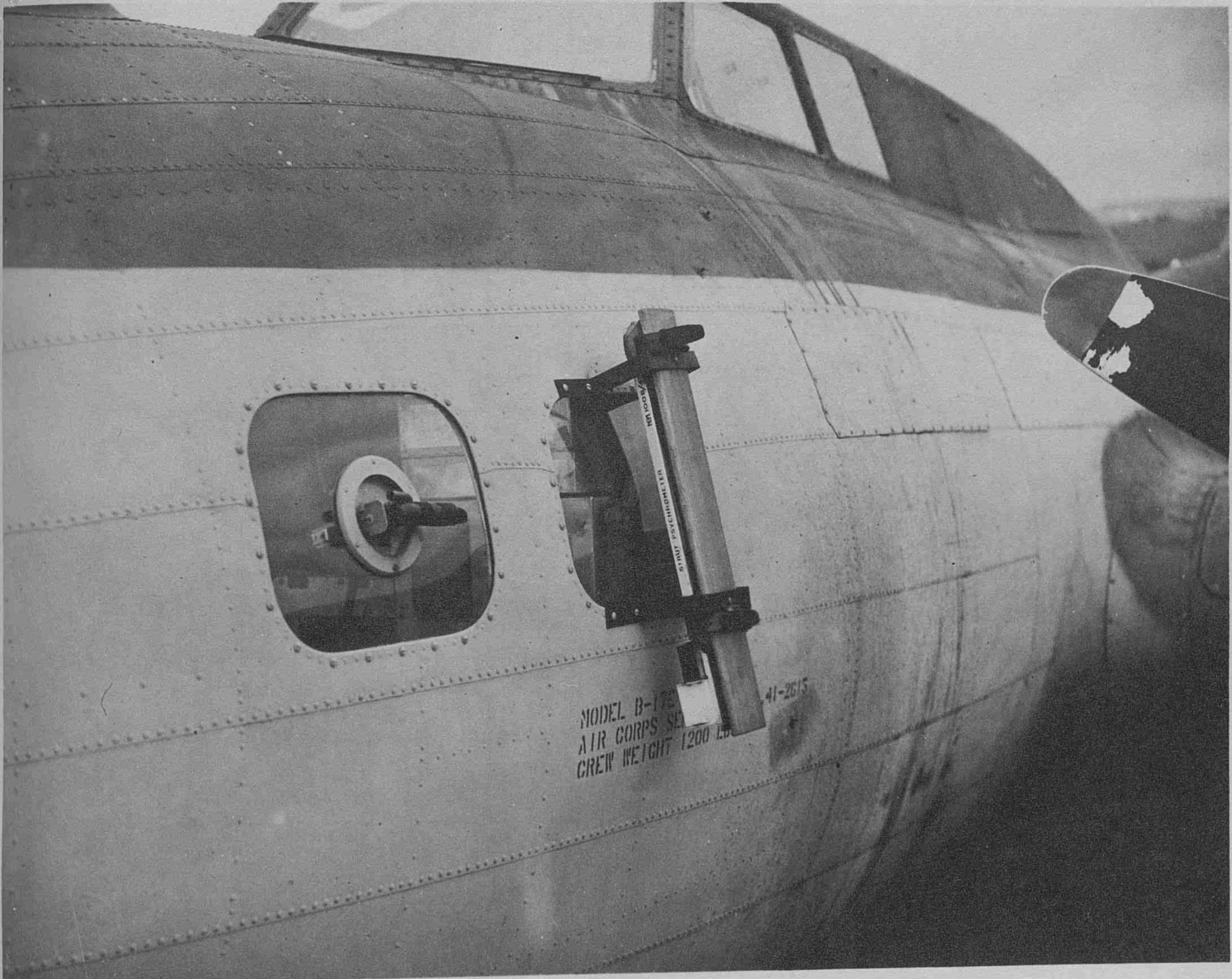

STRUT PSYCHROMETER
MODEL B-17
AIR CORPS SE
CREW WEIGHT 1200 LB
41-2615

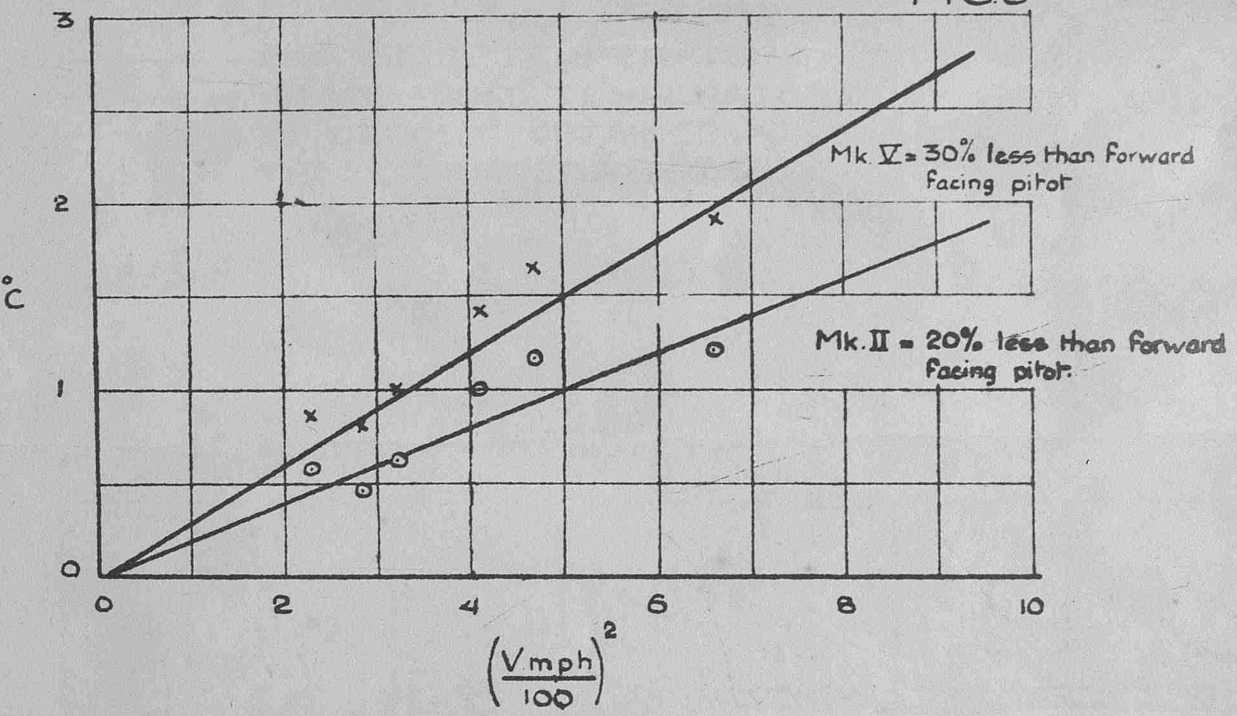
Fig. 2.

SPEED CORRECTIONS TO THERMOMETERS ON FORTRESS F.K 192

⊙ = FORWARD FACING PITOT - Mk.II PSYCHROMETER DRY BULB

X = FORWARD FACING PITOT - Mk V PSYCHROMETER DRY BULB

FIG.3



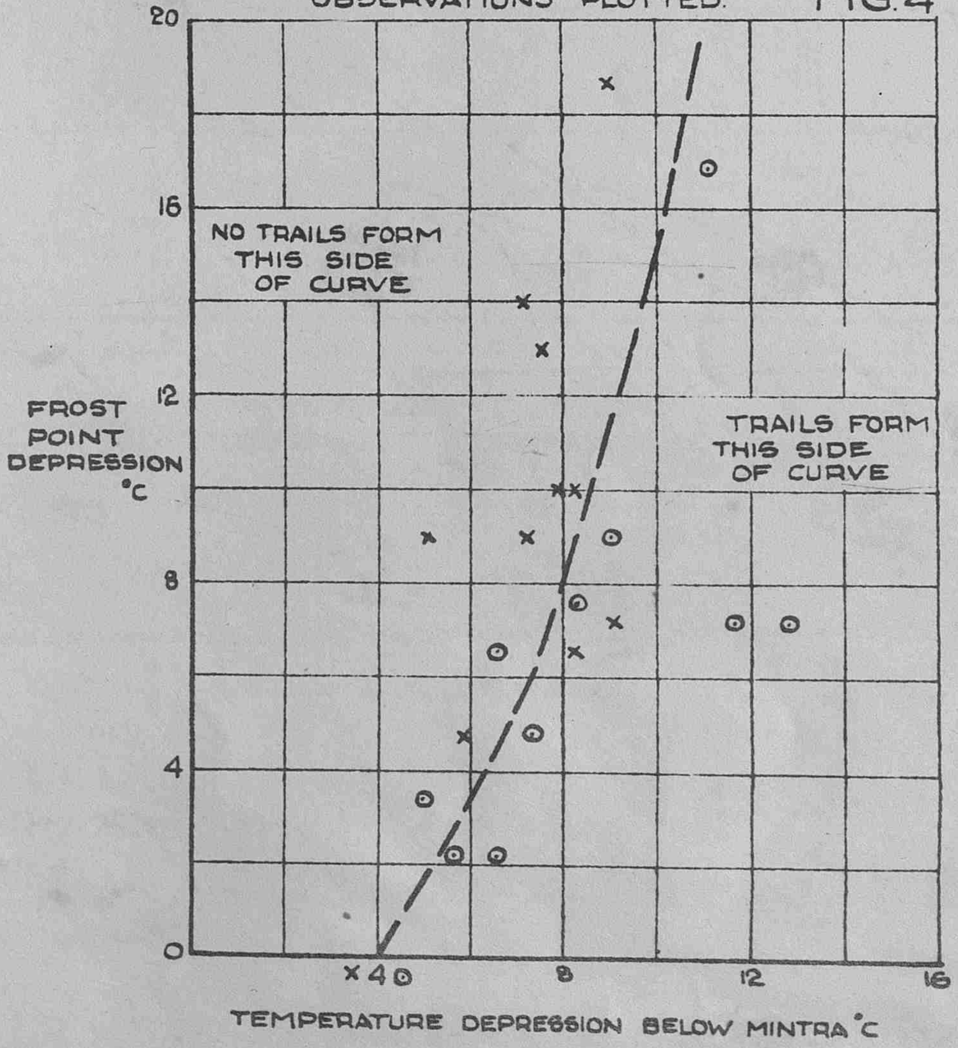
RELATION BETWEEN TRAIL COMMENCEMENT, TEMPERATURE & FROST POINT DEPRESSION.

X = NO TRAIL FORMED

⊙ = TRAILS FORMED.

ONLY BORDER LINE OBSERVATIONS PLOTTED.

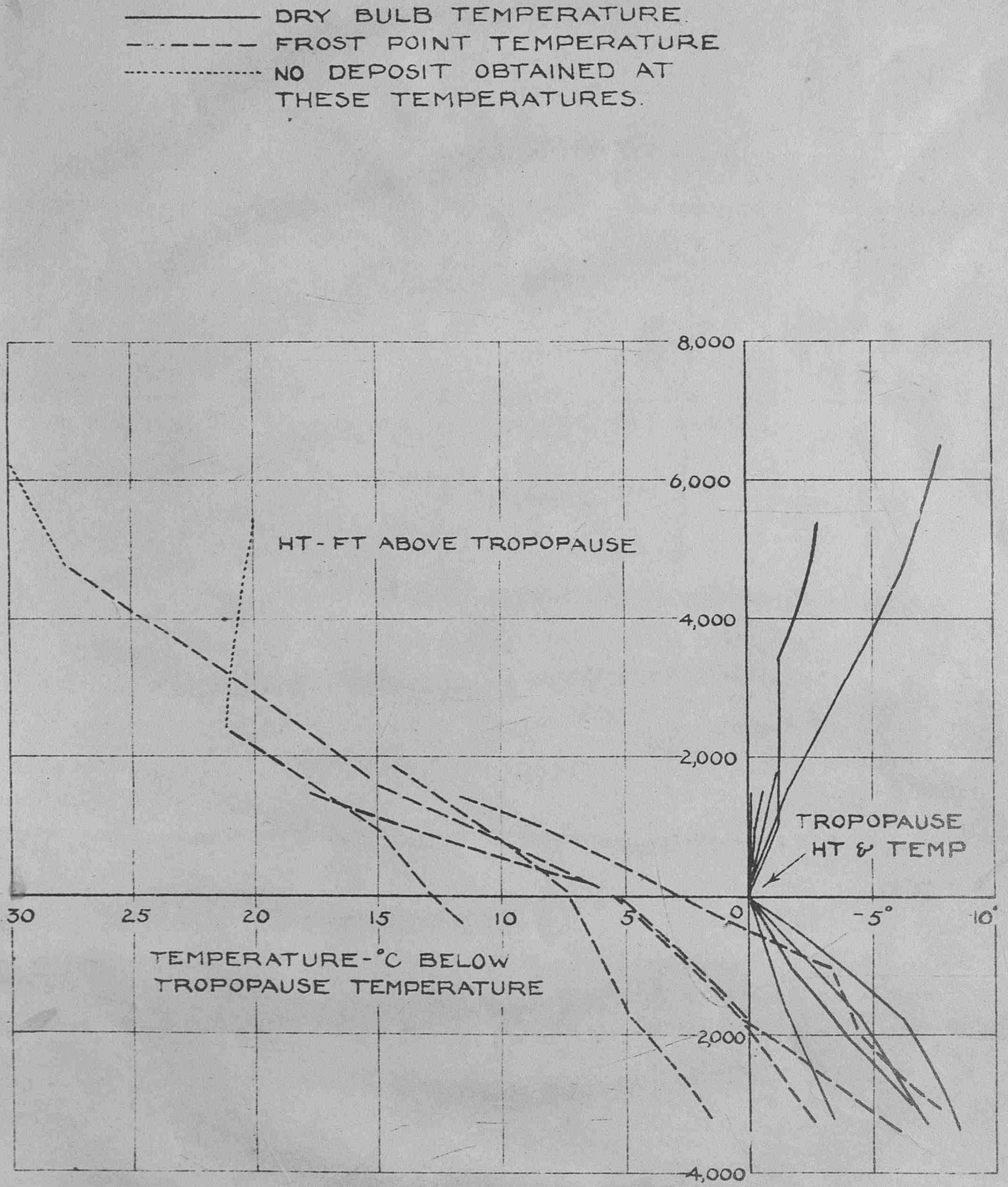
FIG.4



CURVE No 6047

HUMIDITY IN THE STRATOSPERE.

FIG. V



CURVE No. 6048

Form 2811 AIR MINISTRY, METEOROLOGICAL OFFICE DATE 19

TEPHIGRAM OF RADIO-SONDE RECORDS

STATIONS	TIMES

REMARKS

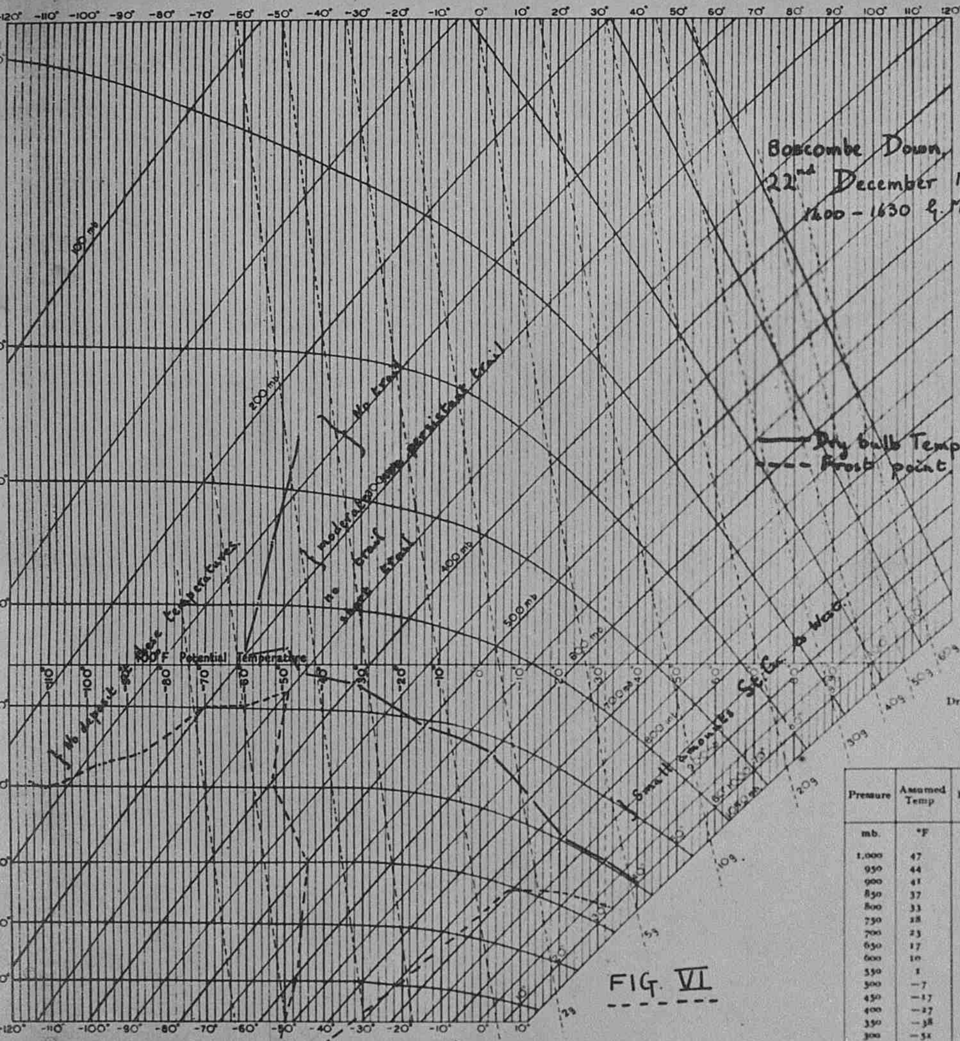


FIG. VI

EXPLANATORY NOTE

This chart is intended for use with radio-sonde ascents to show the general features of the complete results at a glance. It is not intended for use where the main purpose of plotting the result is to examine the details in the lower part of the troposphere. The scale of the form is approximately half that of the Tephigram Form 2810.

The temperature figures printed along the 1,000 mb. line apply both to the saturated adiabatics (the curved lines) and to the temperature scales (the vertical lines). The values printed at the left hand side of the diagram apply only to the saturated adiabatics.

A straight line on the diagram gives a constant lapse rate. For an inversion the curve must cross the temperature scale from left to right. If the result of a radio sonde ascent is plotted it is usually easy to see where the stratosphere begins. It may begin as low as 400 mb.

Dry Adiabatic Lapse Rate is 5.3°F. per 1,000 feet.

Pressure	Assumed Temp	H Height above 1,000 mb	Increase of height due to 10°F increase in mean temperature.
mb.	°F	Feet.	Feet.
1,000	47	0	—
950	44	1,380	30
900	41	2,830	60
850	37	4,350	90
800	33	5,930	120
750	28	7,640	150
700	23	9,420	180
650	17	11,320	230
600	10	13,340	270
550	1	15,490	320
500	−7	17,810	370
450	−17	20,330	430
400	−27	23,080	490
350	−38	26,120	560
300	−51	29,530	640
250	−62	33,440	740
200	−66	38,150	860
150	−66	44,200	1,000
100	−66	52,700	1,200

Adjust heights to other surface pressures by adding (subtracting) 28 feet for every millibar of pressure above (below) 1,000 mb.

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AIR MINISTRY, METEOROLOGICAL OFFICE

DATE...19

TEPHIGRAM OF RADIO-SONDE RECORDS

STATIONS	TIMES

REMARKS

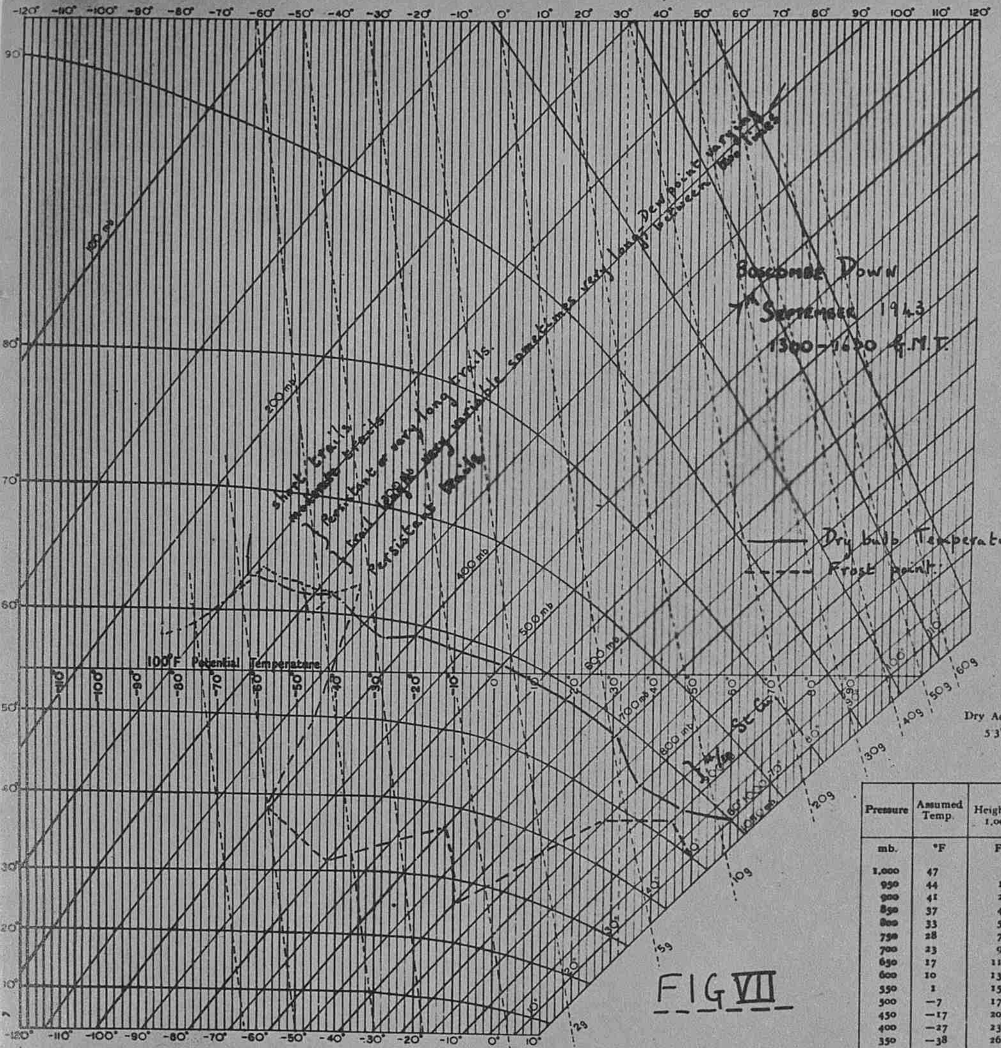

FIG VII

Dry Adiabatic Lapse Rate is 5·3°F. per 1,000 feet.

Pressure	Assumed Temp.	H Height above 1,000 mb.	Increase of height due to 10°F increase in mean temperature.
mb.	°F	Feet.	Feet.
1,000	47	0	—
950	44	1,380	30
900	41	2,830	60
850	37	4,350	90
800	33	5,950	120
750	28	7,640	150
700	23	9,420	190
650	17	11,320	230
600	10	13,340	270
550	1	15,490	320
500	−7	17,810	370
450	−17	20,330	430
400	−27	23,080	490
350	−38	26,120	560
300	−52	29,530	640
250	−62	33,440	740
200	−66	38,150	860
150	−66	44,200	1,000
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Adjust heights to other surface pressures by adding (subtracting) 28 feet for every millibar of pressure above (below) 1,000 mb.

EXPLANATORY NOTE

This chart is intended for use with radio-sonde ascents to show the general features of the complete results at a glance. It is not intended for use where the main purpose of plotting the result is to examine the details in the lower part of the troposphere. The scale of the form is approximately half that of the Tephigram Form 2810.

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A straight line on the diagram gives a constant lapse rate. For an inversion the curve must cross the temperature scale from left to right. If the result of a radio sonde ascent is plotted it is usually easy to see where the stratosphere begins. It may begin as low as 400 mb.

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Icing in Cloud - Analysis of recent reports from aircraft.

By way of supplement to MRP 168 an analysis has been made of all reports of icing as experienced on all the meteorological flights in the British Isles and published in the Upper Air Supplement to the D.W.R. from January 1943 to March 1944 inclusive. The following are the results:-

1. Icing in Relation to Temperature.

Reports are mainly in the form "icing in cloud X mb. to Y mb." and the icing limits stated are practically always the cloud limits or from the freezing level up to the upper limit of the cloud. Probably it is not possible to tell when icing ceases. For this reason the temperature tables are given in two forms (a) the whole range covered by icing and (b) the temperatures at which icing commenced. Table I includes only cases of moderate and heavy icing.

Table I.

Relation of icing to temperature - Percentage frequencies.

(a) All temperatures within the heights at which icing was reported.

°F	-17 to -13	-12 to -8	-7 to -3	-2 to +2	3 to 7	8 to 12	13 to 17	18 to 22	23 to 27	28 to 32	33 to 37	38 to 42
Clear	2	2	3	5	8	13	18	22	20	7	0	-
Rime	2	3	4	5	7	12	17	22	17	7	3	1

(b) Temperatures at which icing commenced.

°F	-17 to -13	-12 to -8	-7 to -3	-2 to +2	3 to 7	8 to 12	13 to 17	18 to 22	23 to 27	28 to 32	33 to 37	38 to 42
Clear	1	-	-	3	1	3	11	16	38	23	3	-
Rime	-	-	6	1	5	8	4	23	24	21	5	4
Both types	1	-	3	2	3	6	7	20	30	22	4	2

From the second table it is seen that at very low temperatures clear ice is less often formed than rime. This is in agreement with the results from mountain observatories (Section 4 of M.R.P.168). The risk of heavy or moderate icing of any type commencing seems to be mainly in the region of 18 to 32°F and few cases occur below 8°F.

2. Icing in Relation to Vertical Development of Cloud.

It is noted that the clouds giving moderate and heavy icing, whether clear or rime, had their bases in about 40% of cases at or below the freezing level and in a further 25% of cases below the 25°F isothermal.

On the other hand clouds giving only light rime or frost or light clear ice had their bases at or below the freezing level in only 10% of the cases and had as a rule relatively small vertical extension. Clouds giving only light icing had in fact an average vertical range of temperature of less than 10°F and in half the cases a range of 5°F or less. Clouds giving heavy or moderate icing had an average range of about 20°F corresponding to a thickness of some 5,000 to 7,000 feet and only one fifth had a range of 5°F or less.

From this it seems fairly definite that the clouds with a serious icing risk are usually those of considerable vertical extension, especially in cases where they begin below or near the freezing level. This is in line with remarks made in

/paragraphs

CLASSIFICATION CHANGED TO ~~RESTRICTED~~ Unclassified
By authority of MRP 385 12 Oct '46
EEH
8-11-52
Date

paragraphs IV(b) and (c) on p.8 of MOM 393 - Note for the Guidance of Forecasters on the Subject of Forecasts of Ice Accretion on Aircraft.

3. Bias in the observations.

The amounts of cloud in cases where icing is reported were as follows:-

Cloud amount	10	9	8	7	6	5	4	3	2	1
% of cases	54	15	13	6	2	6	1	2	0	1

Thus icing is seldom reported in any cloud except in cases where the whole or nearly the whole sky is covered with that type of cloud. The primary duty of pilots being to obtain a record of temperature and weather up to a specified height, they naturally choose a path in reasonably clear air so far as possible and thus, except in cases where a cloud type covers the whole or nearly the whole sky, the tendency is for the interior of that cloud not to figure in the conditions recorded. For this reason the observations treated simply en masse give a biassed picture of the relative liability of the various cloud forms to icing. At first sight (Line (a) of Table II) they suggest that moderate or heavy icing of one kind or another occurs in 10% of all cases of Nimbo-stratus but only in about 2% of cases of Cumulo-nimbus and still less frequently in other cloud types. When however consideration is limited to cases where the cloud covered 8/10 or more of the sky, the relative liability to moderate or heavy icing, as shown in line (c) of Table II, rises to 8% for Cb., 6% for Cu and round about 3% for Sc, Ac, and As., the percentage frequency of icing still remaining highest (10%) in Ns.

Table II - Percentage frequency of heavy or moderate icing either clear or rime.

		St	Cu	Ns	Sc	Cb	Ac	As	Cs	Ci
(a) All amts. of cloud		1.1	1.5	10.1	1.5	2.3	1.2	1.7	0.5	-
(b) Cloud Cover 8/10 or more	(Heavy rime	-	1.0	2.0	0.4	-	0.6	1.0	-	-
	(Moderate rime	1.0	2.7	3.3	1.0	2.8	1.5	0.9	1.0	-
	(Heavy clear	0.5	1.0	3.3	0.3	-	0.3	0.2	-	-
	(Moderate clear	0.5	1.4	1.5	1.1	5.6	1.2	0.7	0.5	-
(c) Cloud Cover 8/10 or more	(Heavy or (Moderate (clear or (rime.	2.0	6.1	10.1	2.8	8.4	3.6	2.8	1.5	-

Separation of the occurrences into heavy and moderate rime and heavy and moderate clear ice - line (b) of Table II - produces the result that, with cloud cover 8/10 or more, Ns leads again with heavy icing whilst Cb shows no heavy icing of either type, but only moderate icing. Indeed when all amounts of cloud cover are considered there are in the whole series only three reports of heavy icing in Cb, namely heavy rime on one occasion with 8/10 Cu and Cb combined, heavy rime on another occasion of 3/10 Cb., and heavy clear ice on one occasion with 2/10 Cb. But for reasons stated above, this by no means excludes the probability that catastrophic icing is a risk in Cb clouds.

4. Comparison with earlier analysis of reports from meteorological flights. In January 1941 a note was prepared in the M.O.9(I)

Branch of the Meteorological Office, comparing icing conditions over Germany and England. It included tables based on the data from the Meteorological Flight at Duxford in Winter and Spring from 1934 - 1938, and from the flights at Berlin, Hamburg and Norderney from 1936 - 1938. In that note also comment is made on the rarity of observations of icing in cumulus and cumulo-nimbus cloud. The following table is a summary derived from Table III of that note:-

Table III. Percentage frequency of icing in different types of cloud.

(Winter and Spring Combined).

		St	Cu	Ns	Sc	Cb	Ac	As	Cs
Duxford	All types of icing.	4.0	1.5	0.5	3.5	0.6	0.5	0.6	0.1
	Heavy ice only.	0.3	0.1	0.1	0.3	0	0.1	0.1	0
Germany	All types of icing	5.4	0.5	3.0	4.4	0.7	2.1	4.0	0.1
	Heavy ice only.	1.6	0.3	1.2	1.2	0.3	0.5	0.8	0

A close comparison cannot be made between Table II and Table III because they are not arranged in the same way, and they do not cover the same seasons of the year. The percentage frequencies seem however to be of the same order except as regards the much higher place taken by Ns according to the more recent observations. This may be because there is now more flying in bad weather, or it may be because at Wick and Aldergrove, which figure in the 1943-44 data, Ns characterized by icing is more often experienced.

5. Summary.

(i) As regards temperature, icing in cloud normally commences between 32°F and 18°F, and very seldom below 8°F; the lower limit to which the icing may continue is less clearly defined. At very low temperatures the liability is to rime rather than clear ice.

(ii) Clouds giving only light icing or frost have as a rule relatively small vertical extension, and seldom begin below the freezing level. Clouds giving moderate or heavy icing frequently have their bases below the freezing level or anyway below the 25°F isothermal, and on the average extend in the vertical direction over a range of temperature of the order of 20°F (some 5,000-7,000 feet).

(iii) As regards type of cloud, the percentage frequency of occurrence of heavy or moderate icing is - according to the statistics - greatest (10%) in nimbo-stratus, but this may be because Ns, when it occurs, usually covers nearly the whole sky. Under the same conditions - almost the whole sky covered - Cb gives a frequency of 8% and Cu about 6%. Bias in the observations is referred to, and the probability that catastrophic icing is a serious risk in Cb clouds is by no means excluded.

F.M.Jones.

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AIR MINISTRY

M.R.P. 185
24th May, 1944.

METEOROLOGICAL RESEARCH COMMITTEE.

The use of salt solution for wetting the Wet Bulb Thermometer.

by

G.C. Simpson.

In 1941 experiments were made at Kew to see whether a solution of common salt could be used for wetting a wet bulb thermometer in order to obtain measurements of the relative humidity at temperatures below freezing point.

It is obvious that a saturated solution would be best for this purpose, not only because the freezing point is lowest, but because it would be difficult to maintain any given degree of concentration other than saturation.

The first experiments were made with Assmann psychrometers. Over the range of natural temperatures and humidities available (by good fortune we had a spell of very dry weather with humidities down to 31%) it was found that the relative humidities recorded by a psychrometer wetted with salt solution (the reduction tables for ordinary water being used) were consistently 25% above those measured at the same time with a psychrometer wetted with water.

The lowest temperature experienced at Kew during the experiments was in the neighbourhood of 40°F. In order to carry the experiments to lower temperatures arrangements were made with the Superintendent, Ditton Laboratories, East Malling, for measurements to be made in the low temperature laboratory similar to those made at Kew. The results showed that steady readings of an Assmann psychrometer could be obtained right down to the freezing point of the salt solution (approximately 0°F) and that the psychrometer wetted with the salt solution recorded humidities 20% above those recorded by one using water. The difference between 25% found at Kew and 20% found at East Malling was due to the different samples of salt used, as with the salt used at East Malling we also obtained a rise in the apparent R.H. of 20%; while later experiments at East Malling with a sample of our salt gave a rise of 25% at humidities round about 45% and a temperature near -8°C; but 20% with R.H. 80-90% and temperatures near -15°C. These slight discrepancies were not finally cleared up before the experiments came to an end. The results however were quite clear that reliable humidities can be obtained by using a saturated salt solution in an Assmann psychrometer down to temperatures near 0°F.

The Assmann psychrometer is used in exactly the same way with salt solution as with water and the relative humidity is obtained by the usual tables; all that is then necessary is to deduct 25% from the relative humidity obtained from the tables. When the humidity is above 75% the bulb wetted with solution reads higher than the dry bulb; but the tables extrapolated above 100% R.H. by the usual formulae still apply. Tables extrapolated for the purpose were computed at Kew.

Experiments were then made to see if salt solution could be used in an ordinary wet and dry bulb hygrometer in which the liquid is fed to the bulb through a wick from a reservoir just below the bulb. Here difficulties were at once encountered; they were of two kinds: (a) when the relative humidity is below 75% and evaporation of the solution takes place the bulb and wick soon become encrusted in salt which ultimately cuts off the supply of solution and the salt encrusted bulb acts as a dry bulb and (b) when the humidity is above 75% water is deposited from the atmosphere on to the wet bulb, and this water flows down the wick so preventing the flow of solution upwards and ultimately removing all the salt from the muslin round the bulb, when the temperature recorded is that of a wet bulb supplied with pure water.

An attempt to get over these difficulties was made by fixing the reservoir of solution above the bulb, so that there was a small flow of solution over the muslin covered bulb, the excess solution falling from the lower part of the bulb at the rate of 5 to 10 drops a minute. This arrangement was successful at Kew when the rate of flow could be watched and adjusted; but careful watching was necessary because if the flow is too small salt appears on the bulb, while if it is too fast the liquid has not time to cool to the full extent and so the temperature recorded is too high. The rate of flow is also affected by the temperature, being twice as rapid for the same head at +10°C as at -10°C.

The investigation had been undertaken with the object of using salt solution with the standard wet and dry bulb hygrometer used on aeroplanes. One of these instruments was obtained and a reservoir attached above the bulb in such a position that the drip was sufficient. Eight flights were made with this instrument with very disappointing results. It was clear that the flow of solution to the bulb was not satisfactory especially at low temperatures probably due to the high viscosity of a salt solution at low temperatures.

As it seemed that further work was not likely to lead very quickly to a solution of the problem, and as another type of hygrometer similar to the one used on radiosondes was being developed, it was decided to go no further with the use of salt solution. My personal opinion is that while satisfactory observations can be obtained with a wet and dry hygrometer down to 0°F by using salt solution, so much care and attention are required that it is not likely to be a useful method for routine observations either on the ground or in the air.

Kew Observatory,
Richmond, Surrey.

23rd May, 1944. (signed) G.C.SIMPSON.

L Gr Brit

AIR MINISTRY

MRP 187
19th August, 1944

METEOROLOGICAL RESEARCH COMMITTEE

THE POSSIBLE USE OF SOLUTIONS OTHER THAN SALT FOR WET BULB THERMOMETERS.

by A. W. Brewer, with comments by Dr. E. J. Bowen

In M.R.P. 185 Sir George Simpson described experiments on the use of common salt as a cryoscopic agent to lower the freezing point of the water wetting a wet bulb thermometer. The conclusion is reached that though under carefully controlled conditions the wet bulb operates satisfactorily, incrustation of the wick by dried salt soon gives trouble and prohibits its routine use. It has therefore been suggested that it may be possible to use a liquid cryoscopic agent which will have the desired effect and yet any surplus solute on the bulb will run away and prevent incrustation.

For a unionised solution the lowering of the freezing point is given by:-

$\frac{18.5\,s}{M}$ (where s is the number of grams of solute in 100 grams of solvent and M is the molecular weight of the solute. Clearly then a high solubility and a low molecular weight are required to attain a reasonable depression of the freezing point.

The liquid must also satisfy the following requirements:-

(a) Must not be miscible with water in all proportions as this would permit drying out all the water leaving the wick covered with pure or nearly pure solute.

(b) Low vapour pressure, (high boiling point) to prevent large additional cooling due to evaporation of the solute. A vapour pressure of about one fifth of that of water would be about the largest permissible.

(c) Low melting point, or else the property of super cooling very greatly (such as glycerine which has a melting point of 20°C. but is rarely seen in the solid state).

(d) Non-corrosive, not dangerously poisonous and reasonably easily obtainable on an adequate scale.

No inorganic substance can be found possessing the required properties.

A list of organic substances some of which may be suitable is given below, together with such data as are available of their properties*. The list is purely tentative and the advice of a chemist seems desirable. When possible the price for 500 grams as listed in a 1940 B.D.H. catalogue is given as a rough index of the commercial availability of the particular substance.

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* Obtained from "The Chemists' Year Book 1941"

SUBSTANCE	MOL.WT	SOLUBILITY	MELTING POINT °C	BOILING POINT °C	PRICE for 500gm.
Acetonyl Urea	123	Very soluble	41	175	not listed
Butylene Glycol (iso or B)	88	-do-	liquid at ordinary temp.	200	Listed but not priced
Chlor Pyridine (2 or 4)	113	Soluble	"	168	Not listed
Hydroxybenzaldehyde	122	"	-21	196	14/6d.
Penta Methylene diamene(cadaverine)	102	"	-34	179	Not listed
Diethylene Glycol	106	"	liquid	25	9/10d.
Lutidine	107	30%	"	142	10/9d.

Of these Butylene Glycol if reasonably obtainable would seem to offer most possibility, as it is given as very soluble and it has the lowest molecular weight. Diethylene glycol is a familiar compound which would appear to be a reasonable choice. The precise data of the solubility required to calculate the depression of the freezing point are only available for Lutidine, which should give a depression of 5°C.

Blocking of the wick with pure solute is still possible even with a liquid, as the wick may not carry it down or the surplus liquid solute may form a film encasing the wet bulb and preventing evaporation. In this connection a mixed wick of muslin and, say, wool might be worth trying. It is well known that water does not wet wool very well but it is possible that the solute might wet it well and the wool could then convey the pure solute to maintain an equilibrium condition on the thermometer bulb surface. This equilibrium under dynamic conditions of rapid evaporation, will also depend upon the diffusion of solute within the solvent to prevent excessive local supersaturation.

It would seem that though the conditions which have to be satisfied are rather exacting, there is reasonable hope that a suitable substance can be found out of the great range of organic substances. The importance, however, of wind tunnel trials in which a considerable amount of evaporation is done is to be emphasised.

Comments by Dr. E. J. Bowen

I entirely agree with paragraph 3. It should be added that the liquid should not be of high viscosity. The 'drying effect', which gives a salt incrustation, will give a similar layer of liquid which will give trouble if not removed. A mobile liquid is essential if this effect is to be minimised. In the note it is suggested that wool might be used to soak up the pure liquid formed by drying out. Another possibility is to try to collect it as drops on some kind of protuberances. A low viscosity is also necessary for ready mixing to take place when the solution is being diluted by taking up aqueous vapour.

The following are some comments on the substances suggested in the note:-

Acetonyl Urea	No solubility data available. Worth considering.
B. Butylene glycol	Very viscous liquid and therefore unsuitable.
3 or 4 chlorpyridine	Probably not soluble enough.
o - hydroxybenzaldehyde (salicyl aldehyde)	Solubility only 1.7 gm/100c.c.
Penta methylene diamene	Viscous liquid, quite unsuitable.

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Diethylene glycol	Viscous liquid. Possibly worth considering. (Not to be confused with ethylene glycol).
2:4 lutidine (2:4 dimethyl pyridine)	Solubility 20 gms. - 100 c.c. Probably the best choice for a start.
2:6 lutidine	is stated to be completely miscible with cold water and to be less soluble in hot. A mixture of 2:4 and 2:6 lutidine would enable variations of solubility to be tried, and a solution to be prepared of such low water vapour pressure that it always took up water from the air. One would then have only the 'dilution' difficulty to overcome and would not be troubled with the 'encrustation' effect.

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ARCTIC CONDITIONS AND EUROPEAN WEATHER

by Gilbert T. Walker

Summary. During the present century a considerable amount of work has been devoted to the relations between ice in the arctic region and European weather; but during the past thirty or forty years the mechanism of control seems to have changed fundamentally and this paper represents a preliminary effort to find out what relations have been holding during recent years.

Some of the anomalies of arctic relationships are traced to 'surges' and the effects of these can be partly got rid of by taking departures from smoothed curves.

An attempt has been made to provide a satisfactory series of annual values of the arctic pack ice that drifts along the Greenland Sea and to correlate it with conditions of pressure and temperature in Europe. Relationships of the ice with previous conditions are closer than those with subsequent ones, and lead to a forecast for which R = .66 a value too small for ordinary use.

For the ice off the Newfoundland Banks the data are more complete but more baffling. The coefficients of - .76 and - .70 of this ice with previous pressure in Iceland and Greenland, based on over 20 years, have become small and positive: some relationships with European weather are found and an attempt to derive an alternative method of forecasting meets with but limited success.

A comparison is made of the results of several methods of smoothing the curves from which the differences are taken, and in an Appendix are discussed some of the effects of these methods upon the correlation coefficients between consecutive terms.

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1. When seeking for explanations of weather changes in Europe one of the most promising suggestions is that conditions may from time to time be disturbed by outbreaks of ice in the northern seas.

The history of weather relationships in the N. Atlantic has been rather surprising, especially of those concerned with the quantity of ice south of Newfoundland. Conditions which were so closely associated for twenty or thirty years that their connexion inevitably seemed to be a physical reality rapidly changed and their relationships greatly diminished or even reversed. While therefore in early years it may have been legitimate to draw inferences from a few years selected for their marked characteristics, this process is no longer safe and statistical methods alone appear permissible. pending clearer physical understanding.

For an admirable and comprehensive account of the literature dealing with arctic ice generally and of the conclusions derived by a number of investigators reference may be made to a recent report by I. I. Schell on 'Polar ice as a factor in seasonal weather' (Monthly Weather Review, Washington, Suppl.39, 27-51, 1940). In view of this report historical questions are here only very briefly handled.

2. Discussions with men of arctic experience, especially J. M. Wordie, have led to the following conclusions:-

(a) Ice estimates made from a point on shore must in general be unreliable unless the observations are made from a height.

(b) The amount of ice depends on the wind direction. In fact the effect of the wind is so great that Speerschneider in his work on the 'State of the ice in the Davis Strait, 1820 - 1930' (p.47), says: "A 'severe ice year' ought to be called a 'year with pronounced or abnormal wind conditions' as the wind is the source while the condition of the ice is the visible effect."

....

(c) We may for present purposes divide the ice cover into 'fast ice' attached in winter to the shore, and moveable 'pack ice'. The fast ice has its greatest width along the Siberian coast, especially near the mouths of the rivers. The extremely heavy pack ice which originates from the Arctic basin and escapes down the Greenland coast is known as the 'Arctic pack' (or Greenland Sea pack ice) and consists of heavy polar floes.

Arctic pack

3. When attempting to form an estimate of the amount of the pack ice it seemed best to use the information collected in the monthly charts and in the text of the 'Isforholdene i de arktiske have', excluding ice in the Kara Sea and the southern edge of the Barents Sea. A discussion of the precision attainable will be found in Appendix A. My estimates were:-

	0	1	2	3	4	5	6	7	8	9
189	.	.	.	.	.	.	.	7	-3	0
190	2	0	4	1	-2	2	1	3	-3	0
191	1	4	3	0	-3	3	3	4	1	0
192	-4	0	-5	0	-1	-6	-4	-3	-1	3
193	-5	-3	-1	-6	-1	-1	-3	-1	-2	.

The primary results of the ice would presumably be upon temperatures, and the following table gives the correlation coefficients between the pack ice and Thorshavn temperature over 40 years. All the coefficients are multiplied by 100 and in view of probable errors of .1 at least, values are given to .02 only. The four seasons December to February, March to May, June to August, September to November are indicated by a, b, c, d:-

Year	a	b	c	d
Two before ice	-18	-12	20	-18
One " "	-34	-02	04	04
Same as "	-44	-18	-32	-30
One after "	-46	-14	-28	00
Two " "	-18	-16	08	-16
Three " "	-38	-36	-02	-10
Four " "	-34	-36	-06	-04

Here the coefficients of the ice with the temperatures in the same year are little greater than those in subsequent years and there is a strange amount of persistence. A similar persistence occurs in the memoir by Brooks and Quennell on 'The Influence of Arctic Ice' (M.O. Geophysical Memoirs 41, 1928) in the effects of Barents Sea ice (Fig.6 and Table XI, pp.16, 17) and of their Ice Index (Fig.14 and Table XXIV, pp.28, 29).

It is hard to believe that abundant pack ice in any single year will of itself produce coefficients of -.34, -.44, -.46, -.18, -.38, -.34 with the winter temperatures of six consecutive years. Moreover a glance at graphs of the quantities in question shows that there is a considerable connexion between the values in successive years. The situation suggests that we may divide the weather changes into three classes. There are variations which are (a) relatively rapid, covering a few days, (b) of medium length, extending over perhaps three to eighteen months, as when we have a mild winter, (c) of slow change, perhaps over four to twelve years.

We are now concerned with (b) and (c), which we may call 'medium' and 'slow' respectively; and as they may have different origins and obey different laws it is advisable to consider them apart.

4. For the 'slow' variations we merely need an effectively smoothed curve, and the 'medium' series follows on subtracting the slow values from the original. If from an original series a, b, c ... we form a smoothed one a', b', c' ... where $a' = (a + b + c + d + e)/5$, $b' = (b + .. + f)/5$, etc., we may find big irregularities; for $b' - a' = (f - a)/5$, so that the difference between two consecutive smoothed terms will at times exceed the standard deviation of the smoothed curve itself. So a second smoothing is carried out, the new series being a", b", c" .. where $a'' = (a' + b' + \ldots + g')/7$, $b'' = (b' + .. + h')/7$, etc. Thus $a'' = \left\{a + 2b + 3c + 4d + 5\,(e + f + g) + 4h + 3i + 2j + k\right\}/35$, and the smoothing

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appears to be satisfactory. As an example the winter and summer data of Thorshavn temperature are, the unit throughout being 0.1°C,

December to February

		0	1	2	3	4	5	6	7	8	9
	189								-2	-2	-2
	190	-2	-2	-2	-3	-2	-2	-3	-2	-2	-2
SLOW	191	-1	-1	-1	-2	-4	-5	-6	-7	-7	-6
	192	-4	-2	1	3	5	6	7	7	8	9
	193	10	10	10	10	10	10	9			
	189								-1	5	2
	190	-3	1	-9	6	5	-1	1	-8	1	10
MEDIUM	191	-13	2	5	5	6	-2	-4	0	-14	-3
	192	-5	9	2	5	-8	0	7	2	-2	0
	193	0	-9	9	-1	11	2	-11			

June to August

		0	1	2	3	4	5	6	7	8	9
	189								5	4	3
	190	3	2	1	0	-1	-2	-3	-3	-3	-3
SLOW	191	-3	-2	-2	-2	-2	-2	-2	-3	-3	-4
	192	-4	-4	-4	-4	-3	-3	-3	-3	-2	-1
	193	0	1	3	5						
	189								-2	-3	10
	190	4	5	-14	-5	5	10	1	-13	0	0
MEDIUM	191	0	4	-4	4	3	-2	0	-1	0	3
	192	2	-4	-6	-3	0	5	10	-2	-10	6
	193	7	-8	-1	14						

The effect of the removal of the 'slow' changes is considerable. Instead of the relations of § 3 above we find:-

Year	a	b	c	d
Two before ice	16	-06	22	-20
One " "	-12	20	00	14
Same as "	-12	-02	-46	-38
One after "	-20	-06	04	12
Two " "	-02	00	18	-08
Three " "	06	-18	16	14
Four " "	08	-14	12	12

These results seem plausible enough if we recognise that the errors due to sampling and to inevitable lack of precision in the ice estimates probably combine to a total of about .15. The coefficients then indicate that the only real effect of much ice is a lowering of the Thorshavn temperature in the summer and autumn of the ice year and perhaps in the winter following.

5. Turning now to the 'slow' variations their graphs are smooth and the number of years between successive maxima and minima quite irregular. If we compare the 'slow' series of Spitsbergen winter temperatures with that of Thorshavn the terms are roughly in the same phase and the coefficient between them is .84. For the spring temperature of Thorshavn the prominent part has roughly the phase of the Spitsbergen winter temperature 4 years later. For the temperatures of the Thorshavn summer and the Spitsbergen winter the difference is about 9 years; so while the coefficient 9 years apart is .86, 4 years apart it is .74, 2 years .60, the same year .32 and 2 years in the other direction .08. Between the Thorshavn autumn and Spitsbergen winter the phase difference is about 6 years and for simultaneous years the coefficient is .58.

The amplitude of the slow component is in general appreciably smaller than that

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of the medium. For the pack ice, actual, the standard deviation is 3.0 in our units; for the medium portion 2.4; and for the slow 1.6.

A statistical investigation of the causal relationships of the slow changes would need far longer series than are now available; but for present purposes such an investigation does not appear to be necessary. If we are trying to forecast any element and an approximation to its medium component has been derived by statistical methods, the slow component can be read off with sufficient accuracy from the smooth series of its slowly changing values.

6. After the method of smoothing described in § 4 had been in use for some time it appeared that while the 'medium' variations so got might be fairly free from surges lasting 12 years or more, those of about 6 years might not be adequately reduced. Successive means of 3 and 5 years were therefore formed instead of those of 5 and 7. At a later stage a further change to 2 and 4 years was adopted, and this also effected some slight reduction in the work of computing. These 'de-smoothing' processes may be designated by (5,7), (3, 5) and (2,4). The relationships obtained by their use are in good agreement and experience alone can show which method is best. A comparison of some of their results will be found in § 18 below.

7. Some of the relationships of the Greenland Sea pack ice with seasonal pressures, using (5,7) means, are given in Table 1, where n is the number of years of data.

TABLE 1

Pack ice and pressure

Station	n	Before ice				Year of ice				After ice			
		a	b	c	d	a	b	c	d	a	b	c	d
Anticosti (St.Lawrence)	31	.	.	22	-30	18	-06	12	26	-10	.	.	.
Jacobshavn	37	.	.	.	.	26	-14	50	-10	-12	.	.	.
Ivigtut	26	.	.	.	.	34	-02	.	.	02	.	.	.
Thorshavn	38	.	.	.	12	54	26	-18	-34	-10	06	-26	-16
Spitsbergen	26	.	.	.	16	52	34	12	00	02	42	14	-36*
Bodo	38	.	.	.	10	38	18	-28	-30	-08	.	.	.
Gjesvaer (N. Norway)	34	.	.	.	02	34+/	10	-22	-06	-06	-06	20	-08
Malye Karmakouli (Nova Zemla)	9	.	.	.	.	36	.	.	.	.	.	.	.
Ust-Zylma (U.S.S.R.)	25	.	.	.	.	46	-06	.	.	.	.	.	.
Irkutsk	47	04	-34	-08	04	-20	14	-20o	00	-36	-04	10oo	-08
Berlin	40	.	.	.	-18	02	06	04=	-12	12	-20	00	-04
Copenhagen	40	.	.	.	-16	16	18	-08	-20	06	-04	00	.
Valentia	37	-12	-06	-18	12	22	24	-28	-22	-12	-04	-06	-32
P. Delgada	38	20	-08	-20	-04	-16	-22	-30	48	-12	04	06	12
Mer. Pr. Diff.	32	16	-18	04	-12	-48	-20	-64	28	20	10	-14	16

* For the following winter, r = .00.
+/ For '97 - '10, r = .00; '11 - '30, r = .56.
o For '84 - '04, - .48; '05 - '30, - .02.
oo For '84 - '04, .62; '05 - '30, - .22.
= For '97 - '16, - .28; '17 - '36, .52.

For Ust-Zylma 'actual' data were used in view of gaps in the series. For the last 'station' the data are Peppler's anomalies of the meridional pressure difference over the N. Atlantic between 30° and 65°N. (Beitr. Geophys. 29, 1931, pp.187-200).

It will be seen that at some stations there is evidence of a marked change of relationship between 1900 and 1910. Accordingly in Table 2 the relationships of the pack ice with pressure at five representative stations are given in more detail, separate values being given for the series before and after December 1st, 1905. The (2,4) means were used. It will be noted that at Paris the winter before the ice

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has a coefficient of -.68 from '97 to '05 (9 years), while from '06 to '38 (33 years) the coefficient is .06. The probable error for 9 years is at least .22.

TABLE 2

Pack ice and pressure

Station	Period	d	Year of ice				Year after ice			
			a	b	c	d	a	b	c	d
Paris	'97-'05	-42	-68	-08	-56	34	32	-52	18	.
	'06-'37	-04	06	10	12	-02	04	06	-10	.
	'97-'37	-14	-16	06	-08	10	12	-04	-02	.
Vienna	'97-'05	-54	-38	-16	-72	-38	46	-12	18	.
	'06-'33	-02	-18	18	30	-06	34	-20	-26	.
	'97-'33	-18	-24	10	06	-16	38	-16	-16	.
Helsinki	'97-'05	02	0	-26	06	-04	24	32	-16	.
	'06-'38	04	26	12	-12	-10	02	02	16	.
	'97-'38	04	18	02	-08	-08	06	10	08	.
Bergen	'97-'05	-30	08	-52	-22	18	14	-04	-56	.
	'06-'26	-08	24	26	-32	-32	02	-22	-04	.
	'97-'26	-14	18	02	-28	-14	06	-16	-22	.
Stykkisholm	'97-'05	16	48	18	64	56	-68	-66	-34	-24
	'06-'37	08	46	26	64	-26	-12	08	-44	-04
	'97-'37	08	46	24	64	-10	-26	-14	-42	00

From Tables 1 and 2 it will be seen that heavy ice years are associated with high pressure in the previous winter in E. Canada, Greenland, Iceland, Spitsbergen, Norway, the Faroes and Finland. For low pressure we have to go south - to the Azores, or Austria, or to Siberia. It has been suggested by Wiese and others that the pack ice is associated with the atmospheric distribution on a large scale and its relations with Siberia are of interest. The correlation coefficients of pack ice and Irkutsk pressure covering the year before and the year after the ice are given in Table 1: those for two, three and four years before and after are:-

4 years before	3 years before	2 years before
10 -02 16 -16	-10 00 02 18	28 -10 -08 -08

2 years after	3 years after	4 years after
02 32† -10 14	-02 -04* -26 -14	30 -04 20 00

† For '85-'10, r = .62; '11-'30, -.06.
* For '86-'10, r = -.42; '11-'30, +.44.

8. Some representative coefficients between the pack ice and temperature will be found in the following table:-

....

TABLE 3

Pack ice and temperature

Station	n	Before ice				Year of ice				After ice			
Godthaab + Ivigtut (actual)	35	.	.	.	.	.	.	44	.	.	.	.	.
Thorshavn	40	-12	20	00	14	-12	-02	-46	-38	-20	-06	04	12
Spitsbergen	26	-16	-10	00	-02	-52	-44	-46	-50	-02	-30	02	06
Gjesvaer	30	10	-04	02	-02	-24	-08	-42	-38	04	08	14	26
Bodo	40	-02	-06	12	-12	-18	-10	-34	-08	24	20	34	30
Bergen	30	.	.	.	02	-52	-18	-30	-24	06	14	06	46*
Berlin	40	.	.	.	-22	-28	-24	00	18	14	-16	06	.
Greenwich	41	.	12	12	-22	-40	-18	08	-04	06	00	-12	.

* For the following quarter the coefficient is .36, and after that .06.

It will be noted that in a heavy ice year the temperature is in general that to be expected when pressure is high over Greenland and Iceland, and low at the Azores and Vienna. The warmth at Bodo in the following year is not strongly marked.

9. The present enquiry was entered on with the idea that from time to time large amounts of pack ice might drift away from the great polar icecap and might dominate the weather of northern and part of central Europe. But it appears (see Tables 1, 2) that in general the relations with pressure conditions are closer before than after the ice season of spring and summer, and that the initiative lies rather with the pressure distribution than with the ice; we are accordingly led, in accordance with the dictum of Speerschneider (§ 2, (b) above), to look to the wind as the deciding factor.

Now the available charts of normal pressure distribution in the arctic seas do not include data of the past twenty years and can only be approximate. But it seems that in winter the trough whose lowest pressure covers Iceland and south Greenland extends to Nova Zemla; in spring the trough, though still existent, is not so deep and in summer it has nearly disappeared (Brunt, Physical and Dynamical Meteorology, 1941, pp.10, 11). So in accordance with § 2 (b) above we should expect a year of abundant pack ice to begin with strong E.N.E. winds in the neighbourhood of Spitsbergen and along the northern side of the trough.

10. In order to test this, Tables 1 and 2 have been utilised to work out the departures in the winter pressure that would be associated with an excess in pack ice equal to the standard deviation of that ice. We find for the pressure departures in millibars:-

Anticosti 0.4, Jacobshavn 0.7, Ivigtut 1.3, Stykkisholm 2.3, Thorshavn 2.3, Spitsbergen 2.7, Malye Karmakouli 1.3, Ust-Zylma 1.6, Gjesvaer 2.1, Bodo 1.3, Helsinki 0.5, Bergen 0.4, Copenhagen 0.5, Valentia 1.1, P. Delgada -0.4, Paris -0.6, Vienna -0.6, Berlin 0.1.

If we take a chart of normal winter pressure and add the above departures to the normal pressures at the places in question we find that in a typical chart before heavy ice the depression centred round Iceland is appreciably weakened - being in our case about 2.3 mb. less deep; and that pressure is raised about 2.7 mb. on the northern margin of the trough which has Spitsbergen and north Norway on its margins. In order that the E.N.E. winds along the northern margin may be materially stronger than usual the rise of pressure along the axis of the trough should be materially less than 2.7 mb.; but on this point we have no information.

11. The coefficients of the pack ice with pressure and temperature at Spitsbergen in the preceding winter and of pressures at that time at Stykkisholm and Thorshavn suggest the possibility of forecasting the ice at the beginning of March. The (5,7) coefficients are, using the data of 1912-37 in all cases:-

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	Ice	Spit.pr.	Styk.pr.	Thor.pr.	Spit.tem.
Spitsb. pr.	40	100	64	34	- 18
Stykk. pr.	52	64	100	94	- 62
Thors. pr.	56	34	94	100	- 54
Spits. tem.	- 52	- 18	- 62	- 54	100

The large coefficient of .94 between pressures at Stykkisholm and Thorshavn makes one of these factors superfluous and we retain Thorshavn as having a slightly closer relation with the ice. We then find between the proportional departures - each divided by its standard deviation - the formula (Ice) = .24(Spits.pr)+.31(Thors.pr) -.31(Spits.temp.) with a joint coefficient R = .66. The probable error is about .15 and R is not big enough to justify an annual prediction.

Newfoundland Ice

12. The arctic region for which the ice data are most complete is that to the south of Newfoundland, and, although much work has been done upon its relations with weather elsewhere, important changes make it necessary to bring statistical examination of these up-to-date as far as possible. Representative relations with pressure (5,7) will be found in Table 4.

TABLE 4

Newfoundland ice and pressure

Station	n	Before ice				Year of ice				After ice			
		a	b	c	d	a	b	c	d	a	b	c	d
Anticosti	41	.	.	.	-20	04	08	04	14	.	.	.	.
Sydney N.S.	45	.	.	.	00	04	14	06	-04	30	-06	-02	-04
Jacobshavn	41	.	.	.	-24	-40	-28	-38	-08	-12	-04	12	12
Ivigtut	43	.	.	-08	-40	-46°	-42	-38	-14	-08	10	-08	-02
Thorshavn	48	.	.	.	22	18	-18	-20	-20	-18	.	.	.
Spitsbergen	26	.	.	.	-02	06	-26	06	16	16	-16	-08	.
Bodo	48	.	.	.	20	16	-22	-26	-08	12	-12	10	18
Copenhagen	49	.	.	.	36	52	02	00	-12	-24	06	-04	12
Berlin	48	.	.	.	44	50/	16	-08	-10	-26	-08	02	-06
P. Delgada	43	.	.	.	-26	-16	26	24	24	12	.	.	.
Mer. Pr. Diff.	45	44	28	-22	26	16	22	24	00	22*	18	22	-08

° From '88 to '10 r = -.70; '10 to '30, r = .04.
/ From '89 to '10, r = .70; '11 to '36, r = .36.
* For '86 to '02, r = -.14; '03 to '30, r = .54

Inasmuch as the chief break in the continuity of the polar relationships (see § 7 above) seems to have occurred between 1900 and 1910 the coefficients for five stations have been worked out for the periods before 1906 and subsequently. They are given in Table 5, in which (2,4) smoothing has in general been adopted.

TABLE 5

Newfoundland ice and pressure over each half of the series of years (2,4)

Station	n	Period	c	d	Year of ice				After ice			
					a	b	c	d	a	b	c	d
Stykkisholm	21	'85-'05	08	-22	-76	-40	-40	30	24	30	26	.
	32	'06-'37	22	-20	22	18	-04	-20	-20	-24	-22	.
	53	'85-'37	16	-22	-18	-04	-18	-06	-02	-04	-02	.
Bergen	20	'86-'05	.	22	-08	-36	-04	-30	36	-20	-02	.
	21	'06-'26	.	26	50	-06	38	-10	-50	-16	-16	.
	41	'86-'26	.	24	22	-18	20	-18	-08	-18	-10	.
Helsinki	20	'86-'05	.	06	-16	-10	-08	-26	56	-10	-04	.
	33	'06-'38	.	14	48	-18	-08	-12	-22	-08	-12	.
	53	'86-'38	.	12	22	-16	-08	-16	12	-08	-10	.
Vienna	20	'86-'05	.	24	76	-26	00	-40	-34	-14	32	.
	28	'06-'33	.	24	26	34	-02	10	-18	-10	-42	.
	48	'86-'33	.	24	46	10	00	-10	-24	-12	-16	.
Paris	20	'86-'05	.	36	66	-02	10	-46	-42	-24	16	44
	33	'06-'38	.	50	32	42	10	-22	-46	-02	-20	-08
	53	'86-'38	.	46	44	28	10	-30	-44	-08	-08	12

13. Probably the most important relationship is that between Newfoundland ice and Stykkisholm pressure. Using the 'actual' figures the coefficient is -.84 for the 18 years '86 to '03; for the next 7 years it is .30; for the next 14 years -.42 and for the 13 from '25 to '37 it is .28. For the entire period it is -.26. If (5,7) smoothing is used these figures become -.84, .56, -.38, .22 and -.24. The coefficients for the half periods '87-'11 and '12-'37 are -.56 and -.04 (actual), or -.54 and .00 (5,7).

The big changes have mainly occurred in relations between ice and other factors, not between pressures and temperatures. The obvious suggestion that they are due to the difficulty of securing accurate data in the last century is vetoed by the consideration that coefficients like -.84 and .76 are not produced by bad or random data.

14. The practical effect of this unsteadiness is by no means negligible. There is widespread belief, based mainly on the work of Brennecke and Meinardus, and embodied in 1925 in a formula of prediction by E. H. Smith, that abundant ice is associated with:

(1) an excess of pressure in Nova Scotia over that in S.W. Greenland (Ivigtut),

and

(2) an excess of pressure at Bergen over that at Stykkisholm.

Now a reference to Table 4 shows that the coefficient of winter pressure at Sydney N.S. with Newfoundland ice is .04 and that the coefficient at Anticosti (in the St. Lawrence) also is negligible. With winter pressure at Ivigtut the coefficient from '88 to '10 was -.70, but from '10 to '30 only .04. Thus neither of the stations in (1) has any value in recent years. Regarding (2) the coefficient at Bergen was not even positive for the period '86-'05 when its excess over Stykkisholm has been regarded as important; and though recently the coefficient has increased to .50, that of Stykkisholm has become .22, so that the difference is scarcely reliable. The use of Copenhagen has been advocated, but from '12 to '36 its coefficient has been insignificant (Table 4).

15. Some of the relations with temperature will be found in Table 6.

TABLE 6

Newfoundland ice and temperature

Station	n	Period	c	d	Year of ice				Year after			
					a	b	c	d	a	b	c	d
GJesvaer	33	'95-'27	.	00	16	24	38	-16	10	30	04	04
Bergen	20	'86-'05	.	.	66	52	04	12	.	-40	04	.
	21	'06-'26	.	.	-02	-18	-38	-18	.	16	-02	.
	41	'86-'26	.	10	38	20	-20	-04	.	-14	00	.
Berlin	21	'85-'05	.	-08	48	60	-06	0	-34	-22	-14	44
	31	'06-'36	.	-26	-60	-22	06	22	40	14	-26	-20
	52	'85-'36	.	-02	-24	04	02	16	-16	02	-22	00
Edmonton (Canada)	20	'86-'05	.	.	.	.	36	.	.	.	.	.
	22	'06-'27	.	.	.	.	18	.	.	.	.	.
	42	'86- 27	.	.	.	.	22	.	.	.	.	.

16. If we were searching for a formula for predicting the Newfoundland ice applicable under present conditions it would be necessary to limit ourselves to coefficients based on the period since 1905, and some of these are promising - such as -.60 with Berlin winter temperature, .50 and .48 with Bergen and Helsinki winter pressures and -.50 with Paris pressure in the previous autumn. These coefficients would have to be recalculated when the recent data become available.

17. Some more general relationships may now be considered. The phrase 'a heavy ice year' in the literature seems at times to imply that both the arctic pack ice and the

....

Newfoundland ice are abundant together; and on the other hand their dissimilar coefficients with Iceland pressure suggests an inverse relationship between them. Actually the coefficient between the arctic pack ice and the Newfoundland ice of the same year is .14; between the pack and the Newfoundland ice of the subsequent year it is -.18, and of the year before -.22.

The fact that the arctic pack and the Newfoundland ice vary directly and inversely with the seesaw of pressure between Iceland and the Azores suggests that we should look at their relations with the North Atlantic Oscillation (N.A.O.), which the author has used as an integration of all the factors, temperatures as well as pressures, which play the most prominent part in the seesaw: this oscillation exists through the year (R. Met. Soc. Memoirs IV, No.36, 1932 and No.39, 1937). We find, as we should expect, that the Newfoundland ice is closely related with the N.A.O. of the preceding winter during the years before 1905, and the (5,7) coefficient from '86 to '05 is .82, which is closer than any of the relations in Table 5. From '06 to '30, r = .12 and for the whole period '86 to '30 r = .48. It is a curious fact that in summer the N.A.O. is almost as closely associated with warmth in S.W. Canada as in N. Norway, and examination shows (Table 6) that Newfoundland ice has a coefficient of .36 with Edmonton for the years '86 to '05.

Between the pack ice and the N.A.O., actual figures, the relationships are:

a	b	c	d
-48	-04	-52	-14

Between the pack ice and the 'arctic ice index' of Brooks and Quennell the correlation coefficient is .54.

18. In order to get an effective comparison of the different methods of desmoothing they were all employed on the coefficients connecting the pack ice with Stykkisholm pressure of the four quarters of the same year. These prove to be:-

	a	b	c	d
Actual	36	10	70	-10
(5, 7)	50	36	60	-26
(3, 5)	46	28	62	-22
(2, 4)	46	26	64	-26

The data used were for the 32 years '06-'37, so that the probable error due to sampling is about .12. In view of the surges we expect real differences between the actual figures and the desmoothed; but between the desmoothed coefficients the agreement is as good as could be expected.

Some other comparisons of actual and desmoothed values may be of interest:-

Pack ice and Valentia pressure

Actual	12	08	-32	-20
(5, 7)	22	24	-28	-22

Pack ice and Thorshavn temperature

Actual	-44	-18	-32	-30
(6, 6)	-12	-02	-46	-38

Pack ice and N.A.O.

Actual	-46	-12	-50	-02
(5, 7)	-40	-16	-50	-08

Pack ice and Mer. Pressure Differences

Actual	-34	-02	-70	22
(5, 7)	-48	-20	-64	28

....

APPENDIX B

On the relationships between successive terms

When we say that a weather factor is subject to surges we mean that there is a tendency for an excess or defect to persist for a time; and if we consider r_1, the correlation coefficient between two successive terms, the appearance of surges means the addition of a positive increment to r_1. Inasmuch as the object of desmoothing is to diminish the surges we should expect it to diminish r_1. Table 7 shows the value of $100r_1$ for the three chief factors using the actual values and the results of the three types of desmoothing.

TABLE 7

Arctic pack ice

	Actual	(5, 7)	(3, 5)	(2, 4)
'97-'06	-36	-38	-52	-54
'06-'39	32	-04	-24	-34
'97-'39	20	-12	-32	-40

Newfoundland ice

	Actual	(5, 7)	(3, 5)	(2, 4)
'88-'06	-10	-24	-38	-44
'06-'39	06	-06	-24	-36
'88-'39	02	-04	-28	-38

Stykkisholm winter pressure

	Actual	(5, 7)	(3, 5)	(2, 4)
'87-'06	-30	-42	-42	-54
'06-'37	00	-24	-40	-54
'87-'37	-12	-32	-42	-54

We may note that:-

(a) The values of r_1 form a diminishing series as we pass along from 'actual' to (2, 4) in successively more potent modes of desmoothing.

(b) As the three coefficients from actual figures for the whole periods we have .20, .02 and -.12, suggesting that Stykkisholm is the least affected and the pack ice the most affected by surges.

(c) It is natural to look here for some evidence of the unsteadiness - the change of character in the Newfoundland ice between the first and second periods; the period '97-'05 of the pack ice is too short to give reliable results. If we express the change in the actual Newfoundland r_1 from -.10 to .06 by saying that the effect of the surges has increased by .16, then the effect is still as great as .14 and .08 after the desmoothings (3, 5) and (2, 4). If we compare this with Stykkisholm we find that there the effect has increased by .30 in the actual figures, while in the desmoothed the increase has been only .02 and .00.

It is conceivable that this difference of behaviour of the ice and the pressure series may be associated with the unsteadiness of the former; but the physical origin of the unsteadiness is very obscure.

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

Amendments to M.R.P. 196

Arctic Conditions and European Weather

by

G.T. Walker.

Page 1. §1. In the second section the last sentence should run:-

this process is no longer safe and, pending clearer physical understanding, statistical methods alone appear permissible. ✓

Page 2 §3, line 4. After "Barents Sea" insert "A discussion of the precision attainable will be found in Appendix A". ✓

Page 10. The present appendix to become Appendix B. ✓
Add new appendix A attached.

Stykkisholm pressure				
Original ice series	.38	.14	.78	-.10
Second " "	.26	.06	.68	-.04
Greenwich temperatures				
Original ice series	-.32	-.36	-.16	-.08
Second " "	-.34	-.42	-.34	-.28

The twelve differences between the coefficients have an average numerical value of 0.09: the "probable" value of a difference between two coefficients is roughly .07 - half the differences exceed this amount - so that the probable error of a single coefficient due to inaccuracy of estimation is not more than .06. The probable error of the correlation coefficients due to "Sampling" is about .11 and that due to both causes together about .13. The errors in estimation have therefore not seriously increased the inevitable error due to the shortness of the

/available

AIR MINISTRY MRP 196

METEOROLOGICAL RESEARCH COMMITTEE

Arctic conditions and European Weather.

The accuracy of the ice estimates.

Appendix A.

For the arctic pack ice the area examined has been the Barents Sea (excluding its southerly and easterly margins where land ice forms), the sea round Spitsbergen, and the Greenland Sea. The authoritative Isforholdene contain a descriptive summary of the ice conditions as well as monthly charts showing the ice limits. In the summary the extent to which conditions are treated as "favourable" is that to which the ice is penetrable by shipping. The estimate formed will not be necessarily identical with that derived from the charts, from which a comparison with the normal limits published since 1920 affords a fair idea of the variations of the ice-containing area. For theoretical purposes an area entirely covered by thick ice would have more effect than an area partly covered by thin ice, so that a knowledge of general penetrability, not confined to a few routes, should be important: the ice limits indicated on the charts may be less dependent on accidental local conditions.

The process followed has been to allot estimates 0, ±1, ±2...±7 for the departures at the amounts of ice in the three areas, 2 indicating a moderate excess, 4 a large excess and 6 a very large excess, these estimates being based both on the summaries and on the charts: (the mean of the three estimates was adopted.) In the early years information was scanty and the estimates are somewhat uncertain.

In order to form an idea of the precision in the estimates of §3 I have made new estimates without reference to the former and find for the difference regardless of sign an average of 1.3, with a standard deviation of 1.6. For an indication of the effect of this difference on the correlation coefficients I have work worked out twelve cases - with Berlin and Stykkisholm pressures (actual) for the four quarters of the year of the ice during '99 to '38, and similarly with Greenwich temperature. The values are:-

	a	b	c	d
Berlin pressure				
Original ice series	-.08	.10	.10	.00
Second " "	-.06	.16	.00	-.04
Stykkisholm pressure				
Original ice series	.38	.14	.78	-.10
Second " "	.26	.06	.68	-.04
Greenwich temperatures				
Original ice series	-.32	-.36	-.16	-.08
Second " "	-.34	-.42	-.34	-.28

The twelve differences between the coefficients have an average numerical value of 0.09: the "probable" value of a difference between two coefficients is roughly .07 - half the differences exceed this amount - so that the probable error of a single coefficient due to inaccuracy of estimation is not more than .06. The probable error of the correlation coefficients due to "Sampling" is about .11 and that due to both causes together about .13. The errors in estimation have therefore not seriously increased the inevitable error due to the shortness of the available series of data.

2. On the other hand for the quantity of ice off Newfoundland, where the standard data are those of the American Coast Guard, the figures used are directly based on the actual number of bergs observed. The area in question is well patrolled and the data may be accepted as reliable.

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Comparison of Heights measured by Radio-sonde and by G.L.III

by

D.N.HARRISON

SUMMARY

Heights derived by the standard formula from pressures and temperatures measured with Kew Mark I radio-sondes have been compared with those derived from radar observations on a target carried by the same balloon. The radar observations are sufficiently accurate to give a satisfactory measure of the radio-sonde height errors.

The main results are:

(i) In the first 9 Km the radio-sonde heights are systematically too low, with a maximum error of about 0.1 Km between 4 Km and 6 Km.

(ii) Above 9 Km, the error increases rapidly; positive and negative errors are about equally common, but for any one ascent the sign is constant in the upper levels, i.e. the height becomes increasingly too large or too small.

(iii) There is no evidence of a significant systematic error in the stratosphere. At 20 Km the mean error is only -0,1 Km (equivalent to about 1 mb.) but the mean deviation from this is 0.8 Km (7 mb.)

The question is considered, to what extent the height errors may be attributed to imperfections in the method of correcting the pressure readings for the effect of low temperatures on the pressure unit of the transmitter. Evidence for a lag in the temperature of the P unit as the balloon rises is found in a sharp change of height error at the tropopause.

The effect on the heights of errors in temperature due to solar radiation is also discussed.

Observations on the descent confirm the effect already known, viz. that in the stratosphere pressures tend to be measured lower on ascent than on descent.

I Method and data.

1. Two series of trials have been made at Larkhill, in which the radio-sonde balloon was followed by Radar, the first series in June 1943, using special G.L.III equipment lent by A.D.R.D.E., the second in March-June 1944, with a standard G.L.III, or A.A. No.3 Mark II. These trials have provided data on the accuracy of the radio-sonde observations. It is the errors in the heights computed from the pressure and temperature observations that are considered in this report; the references are generally to the results of the 1944 series of trials.

/2.

2. Most of the ascents of the first series and all those of the second were at about 12 h. G.M.T. The R.S. apparatus and observational procedure were normal, except that a 3 foot sectional target was attached immediately beneath the balloon and the free lift was increased to give an approximately normal rate of ascent. The G.L. readings of slant range r, elevation E and azimuth were made every minute, synchronization errors being negligible. In moderate winds the observations were limited by the scale of the G.L. (r = 36,000 yards or 33 Km), but in light winds the ascent was followed up to the bursting of the balloon, and in some cases readings were made on the descent.

Theodolite observations were made on a number of occasions for comparison with the G.L. readings, from which it was found that the probable error of a single reading of elevation by the G.L. is 0.1°. The resulting random errors in the heights have been largely removed in the computation process; it is considered that any errors remaining are small compared with those of the RS heights, and the height differences (RS - GL) are treated as RS height errors.

3. The method of computation was as follows:-

G.L. III. The height r sin E was computed for every minute, and was corrected for the distance of the transmitter below the target. No correction was applied for the earth's curvature, which at the extreme horizontal range reached is less than 0.08 Km.

Radio Sonde. The normal procedure was followed, except that a larger number of pressure points was computed, usually one per minute, and the amount of smoothing of the " "rough plot" was reduced to a minimum. The height above station level was computed by the standard method in geodynamic units, and converted to kilometres by the addition of 2.0%. This gives heights which are slightly too small in the upper levels, but the error is only -.02 Km at 15 Km and -.05 Km at 21 Km, and has been ignored. The correction for humidity has also been ignored; this arises in the lowest levels, but for a saturated atmosphere and at summer temperatures the correction does not amount to more than about +0.02 Km.

4. The height differences (RS - GL) were plotted against the GL height, a smooth curve was drawn through the points, and the differences were read off for every kilometre of height. Tables I (1943 series) and II (1944 series) give the mean differences in kilometres, with the approximate pressure equivalents, also the "mean deviation", or mean departure from the mean without regard to sign, and the "arithmetic mean", or mean height difference without regard to sign. In Table II the mean height of the tropopause, where reached by the observations is given: this is the true height at the point of inversion. Tables III and IV give the mean differences of RS height error between ascent and descent, for those occasions when the descent was observed. The RS heights for the descent were computed downward from the bursting point, using the temperatures recorded on the descent; the differences given are therefore the height equivalents of the differences of pressure and temperature recorded at given heights on ascent and on descent. The mean difference in pressure at the tropopause between ascent and descent, with its height equivalent, is also given in Table IV: these figures are independent of the GL observations.

/5.

5. Since reference is made in the discussion to the corrections applied to the pressure readings for the effect of temperature, the following outline of the method is given.

The pressure units of the transmitters are calibrated at room temperature. In flight, however, much lower temperatures are encountered, which cause an increase in the frequency signalled at a given pressure. The correction applied is the product fQ, where Q is read from a set of standard graphs as a function of the observed pressure-frequency and the temperature, and f is a constant factor peculiar to the transmitter.

The factor f is measured during calibration, by noting the change of frequency caused by a given reduction of temperature at three pressures, 300 mb., 200 mb. and 100 mb., and comparing these with the corresponding values of Q; three ratios, f3, f2 and f1, are thus found, and the value adopted is the mean of these, rounded off to the nearest .05. Thus f is the ratio of the temperature coefficient of the pressure unit to the mean of the sample from which the standard graphs were derived. It is assumed that this value of f applies over the whole range of pressures.

The Q graphs are revised from time to time. During the trials of 1943 the graphs used were those known as "Q1" and "Q3"; "Q3" graphs were used throughout the 1944 series.

The radio-sonde technique is described in "Meteorological Office Radio-sonde, Measurement of Temperature and Humidity (Radio-thum)", M.O.462.

II. Discussion of data.

1. The most obvious result is the systematic negative error of the RS heights in the lower levels. This suggests that the Q3 corrections are no longer strictly applicable. If this be the explanation, the error ought to be proportional to the transmitter factor. Fig.1 shows the error at 4 Km. plotted against the factor, for all ascents on which Q3 corrections were used. In spite of much scatter there is some support for this theory, though there would not be were it not for the two points in the top left-hand corner of the diagram. There is also a suggestion of a positive height error for zero factor, which would point to an additional source of error. It may also be noted that four ascents of the 1943 series on which Q1 corrections were used show negative errors slightly, though probably not significantly, larger than those under discussion.

The mean error of -0.09 Km at 4 Km. corresponds to an error of +1.4 c/s at about 910 c/s. Such an error could be caused by an error of the B.T.O. scale in a region not corrected by calibration. This, however, will hardly account for the errors found, since the normal calibration point at 900 c/s usually occurs between the heights of 4 Km and 5Km.

2. The factors, f3, f2 and f1 are in many cases not all equal. This means that the Q correction graph cannot be made to suit each individual transmitter by a uniform change of scale. The use of a constant factor f will therefore lead to wrong pressure corrections. Since f3, f2 and f1 usually progress in one direction (increasing or decreasing) f is approximately equal to f2. Thus the effect will be small at 200 mb. and it can easily be shown that, when measured in kilometres, it will be numerically largest at 100 mb: taking average values of frequency and temperature, a change of +0.1 in the

factor will lead to a change of +40 metres at 300 mb and +170 metres at 100 mb. To test this, the height error at 16 Km has been plotted against f - f1(Fig.2); the calculated relationship for 100 mb is shown by a straight line. It will be seen that there is a relationship of the right sign, but larger than that calculated. These facts are further illustrated by the following Table V, which shows the height errors at 16 Km (Δh), and those derived from pressures re-computed with factor f1 instead of f (Δh_1), for seventeen ascents.

TABLE V.

Height errors at 16 Km for factor f1.

Ascent No.	6	9	12	15	17	20	21	24	25	26	27
Δh Km.	-.23	+.37	+.37	-.19	-.24	+.52	-.93	-.55	-.33	+.20	+.71
Δh_1 Km.	-.02	+.34	+.35	-.16	-.18	+.23	-.73	-.52	-.24	+.19	+.54

No.	28	29	30	31	32	37	Mean.	Mean dev.	Arith. mean	Mean (Omitting 31 and 37)	Mean dev. (Omitting 31 and 37)	Arith. mean (Omitting 31 and 37)
Δh	-.64	-.45	+.16	-.26	(+.19)	-.16	-.09	.37	.38	-.07	.40	.41
Δh_1	-.56	-.44	+.21	-.26	(+.10)	-.31	-.09	.31	.32	-.06	.32	.32

In all but two cases (nos. 31 and 37) the error is reduced numerically; these two anomalous ascents are the only ones up to the time of writing of which the transmitter has been found on return to have a leaking aneroid, but this defect does not account for the anomaly, since, if present during flight, it would have given rise to positive height errors. The algebraic mean error is not altered by the correction, but the mean deviation and arithmetic mean error are slightly reduced. Only a small part of the error is, however removed, and Δh_1 still shows a marked correlation with (f - f1); this is unexplained.

The errors at 9 Km show no correlation whatever with (f - f3).

3. The maximum mean (positive) error occurs at about the mean height of the tropopause. This effect is seen more clearly if the measurements are centred at the tropopause. The height errors have been re-measured at kilometre intervals from the tropopause, for eighteen ascents at which there was an appreciable inversion and for which data are available for a further 4 Km. The result is shown in Table VI.

Table VI - Height errors centred at tropopause (18 ascents)

Distance from trop. Km.	-4	-3	-2	-1	$-\frac{1}{2}$	0	$+\frac{1}{2}$	+1	+2	+3	+4
Mean Height error Km.	-.08	-.07	-.04	-.01	.00	+.02	+.02	-.04	-.07	-.09	-.12

The maximum is here sharper than that of Table II, and occurs at ~~within the first $\frac{1}{2}$ Km above~~ the tropopause. Except at the tropopause itself, the height errors found depend to some extent on the method of smoothing (some smoothing is necessary in order to interpolate values at the required points). The measurements from which Table VI was derived were made from the curves referred to in section I 4, with the addition in each case of a point at the tropopause, obtained by comparing the RS height computed from a single pressure reading with an interpolated value of the GL height. The error given for the tropopause is the unsmoothed mean.

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The form of the change at the tropopause suggests that the temperature of the P unit lags behind that of the T unit. At this height, for a factor of 1.0, an error of +10°F in the temperature assumed for the P unit would give rise to an error in the RS height of -~~1~~0.10 Km, which is about the extent of the variation shown in Table VI. Inversions of this amount within the first kilometre above the tropopause are common among the ascents under discussion. The question is complicated, owing to the probable differences in rate of thermal response between the aneroid capsule and the inductance, but qualitatively it can be seen that if the inductance lags more than the aneroid the error will be increased, since the effect of a change of temperature is of opposite signs in the two components. This question is considered further in section II 5.

4. It is of interest to note that there is no change in the rate of ascent at the tropopause. Table VII shows the mean height differences in successive minutes, computed from the unsmoothed GL observations for eighteen ascents (of which sixteen are the same as those used for Table VI). Minute 0 is the minute interval containing the tropopause.

Table VII - Rate of ascent through tropopause (18 ascents)

Minute	-6	-5	-4	-3	-2	-1	0	+1	+2	+3	+4	+5	+6
Mean rate of ascent (m/min)	440	446	463	462	481	502	503	493	527	530	547	520	526

The rate of ascent increases with height. The slight scatter is due to errors of observation, and is not significant. The mean rate of ascent from surface to tropopause for the same ascents is 382 m/min.

This result may not be characteristic of normal radio-sonde ascents, since it may be affected by the increased total lift of the balloon and the drag of the target. A sharp decrease in rate of ascent is usually found at the tropopause, but part at least of this must arise from the errors shown in Table VI; with a constant rate of ascent of 500 m/min, these will cause an apparent decrease of 30 to 50 m/min, depending on the time interval taken.

5. In this section we consider the effect of temperature errors on the RS heights. This enters in two ways, through the layer thicknesses, computed for given pressure levels, and through the Q corrections applied to the pressure observations.

Table VIII shows in the second column the mean difference ΔT between the temperature recorded at midday and midnight at given pressures, for the four months March - June 1944. It is not certain that the whole of these differences represents errors in the day temperatures, but for the present purpose it will be assumed that this is the case. The third column gives the resulting error Δh_T in the height computed for the given pressures, and the fourth column the height error Δh_p due to the error in the Q3 correction for a transmitter with factor 1. Δh_T and Δh_p are of opposite signs, and their sum is the total height error. The last column of the Table VIII shows the true heights at the given pressures.

The total height error is negligible up to 150 mb (13½ Km),

/but

increase of scale is an improvement. Blunders of 5 or 10 c/s or 1 minute can frequently be detected, but not always. Careful joining up of the recorded points is as important as accurate plotting; it should be remembered that each point represents an area ½ c/s x 0.1 minute, within which the true position lies, and very little further smoothing should be applied.

(d) Mistakes may be made in reading from the rough plot or the calibration curves, but these will be less likely to occur if the drawing is well done.

(e) There is a tendency to forget the correction to Q due to the difference of the calibration temperature from standard. This is a very small error, but it would be avoided either by calibrating always at the same temperature or by applying the correction when drawing the original calibration curve.

8. I am glad to acknowledge the great help given to me in the preparation of this report by Flying Officer J. Piegza, of the Polish Air Force, by whom a large part of the work, both observing and computing, has been done.

Table I - Mean height differences (RS - GL) 1943.

Height Km.	Mean diff. Km.	Mean diff. mb.	No. of ascents	Mean Deviation Km.	Arithmetic Mean Km.	Arithmetic Mean mb.
1	-.01	+1	7	.04	.04	4
2	-.06	+6	10	.06	.08	8
3	-.09	+9	10	.07	.10	9
4	-.11	+11	10	.07	.11	9
5	-.08	+8	11	.09	.10	7
6	-.05	+5	11	.09	.08	5
7	-.03	+3	11	.10	.11	6
8	-.01	+1	10	.10	.10	5
9	.00	0	9	.10	.10	5
10	+.02	-2	9	.15	.14	6
11	+.04	-4	8	.19	.17	6
12	+.02	-2	8	.25	.24	7
13	-.03	+3	8	.30	.32	9
14	-.19	+4	6	.19	.28	6
15	-.15	+3	5	.18	.24	5
16	-.15	+3	4	.29	.25	4
17	-.03	0	4	.39	.41	6
18	+.25	-3	2	.50	.50	6
19	+.50	-5	1	-	-	
20	+.22	-2	1	-	-	

/Table II.

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Table II Mean height differences (RS-GL): 1944.

Height Km.	Mean diff. Km	Mean diff. mb	No. of ascents.	Mean deviation Km.	Arithmetic Mean Km.	Arithmetic Mean mb.
0.5	-.01	+3	33	.03	.03	3
1	-.03	+3	41	.04	.04	5
2	-.04	+4	43	.04	.06	5
3	-.08	+7	43	.04	.08	7
4	-.09	+7	43	.05	.09	7
5	-.09	+6	42	.06	.09	7
6	-.08	+5	40	.07	.09	6
7	-.07	+4	40	.08	.10	6
8	-.06	+3	37	.10	.11	6
9	-.03	+2	34	.12	.13	6
10	-.01	0	30	.14	.14	6
11	+.01	0	27	.15	.16	6
12	+.02	-1	23	.20	.20	6
13	.00	0	21	.22	.22	6
14	-.05	+1	20	.25	.26	6
15	-.08	+2	18	.30	.32	6
16	-.09	+1	17	.37	.38	6
17	-.02	0	12	.46	.46	7
18	-.11	+1	10	.50	.52	6
19	-.11	+1	10	.58	.60	6
20	-.07	+1	9	.76	.77	7
21	-.15	+1	6	.96	.96	8
22	-.20	+1	4	1.31	1.31	9
23	+.82	-5	1	-	-	-
Mean trop. Height	11.2	-	22	-	-	-

Table III.

Mean differences (Ascent - Descent)
1943

Height Km.	Mean diff. Km.	Mean diff. mb	No. of ascents.
1	+.06	-7	1
2	-.02	+2	1
3	-.06	+5	1
4	-.05	+4	1
5	+.02	-1	1
6	+.10	-7	1
7	+.09	-5	2
8	+.17	-9	2
9	+.19	-9	2
10	+.23	-10	2
11	+.29	-10	2
12	+.31	-10	3
13	+.30	-8	3
14	+.22	-5	1
15	+.03	-1	2
16	-.13	+2	2
17	+.13	-2	2
18	+.40	-5	1
19	+.34	-4	1
20	+.09	-1	1

Table IV

Mean differences (Ascent - Descent)
1944

Height Km.	Mean diff. Km.	Mean diff. mb	No. of ascents.
1	-.22	+24	1
2	-.05	+5	3
3	+.03	-3	3
4	-.02	+1	3
5	-.04	+3	4
6	-.04	+3	4
7	-.03	+2	4
8	-.01	+1	4
9	+.01	0	6
10	+.08	-3	9
11	+.16	-6	9
12	+.11	-3	9
13	+.08	-2	8
14	+.09	-2	10
15	+.11	-2	10
16	+.16	-3	8
17	+.13	-2	8
18	+.14	-2	8
19	+.13	-1	8
20	+.12	-1	7
21	+.02	0	4
22	-.06	0	3
23	+.06	0	1
Tropopause	+.35	-10	10

Table VIII - Effects of Temperature errors on RS heights.

p mb	ΔT °F (+)	Δh_T metres (+)	Δh_p metres (-)	$\Delta h_T + \Delta h_p$ Km	h Km
40	14	243	768	-.53	22.0
50	12	195	549	-.35	20.5
60	11	161	420	-.26	19.3
70	10	135	329	-.19	18.3
80	9	115	261	-.15	17.5
90	8	100	210	-.11	16.7
100	7	88	169	-.08	16.1
150	4	54	60	-.01	13.5
200	3	38	35	.00	11.7
250	3	28	23	+.01	10.2
300	3	21	16	+.01	9.1
400	2	11	7	.00	7.1
500	1	6	0	+.01	5.5
600	1	3	0	.00	4.1
700	1	1	0	.00	2.9

ΔT = mean temperature difference (12h-24h), March-June 1944.

Δh_T = height error at given pressure due to temperature error ΔT.

Δh_p = height equivalent of pressure error due to ΔT (Q3 corrections, f = 1).

h = height at true pressure p with night temperatures

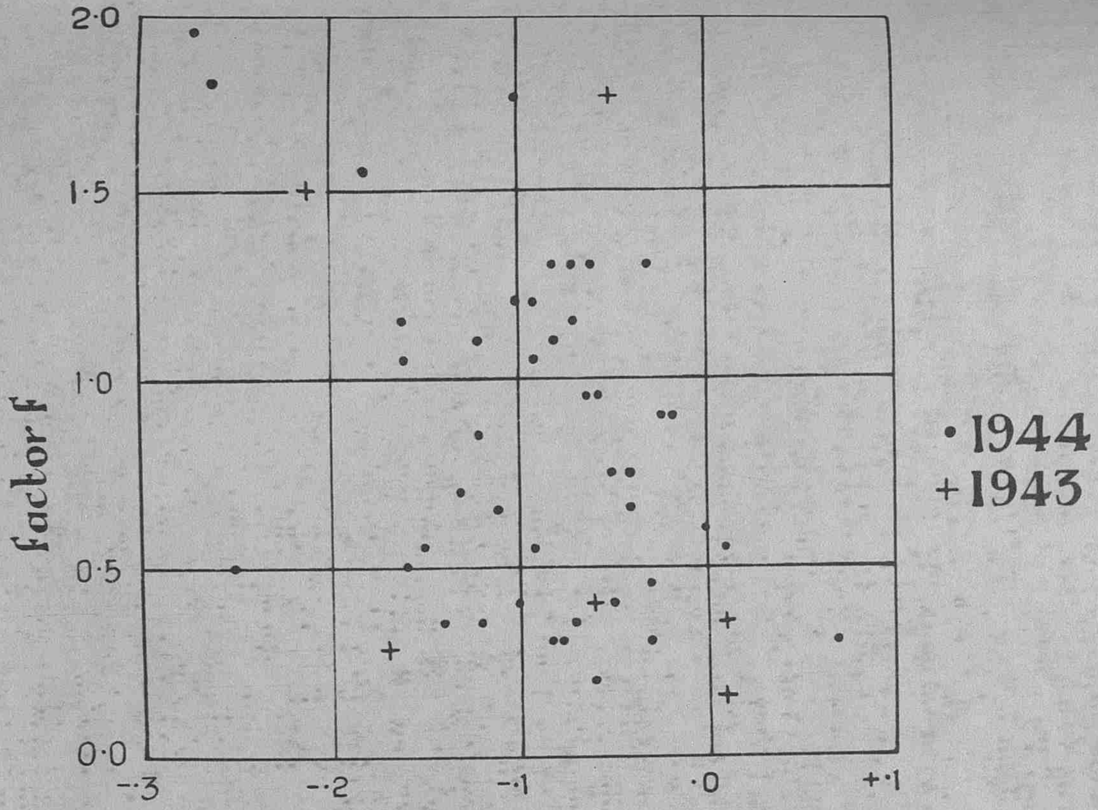

Fig. 1. Error at 4 Km.

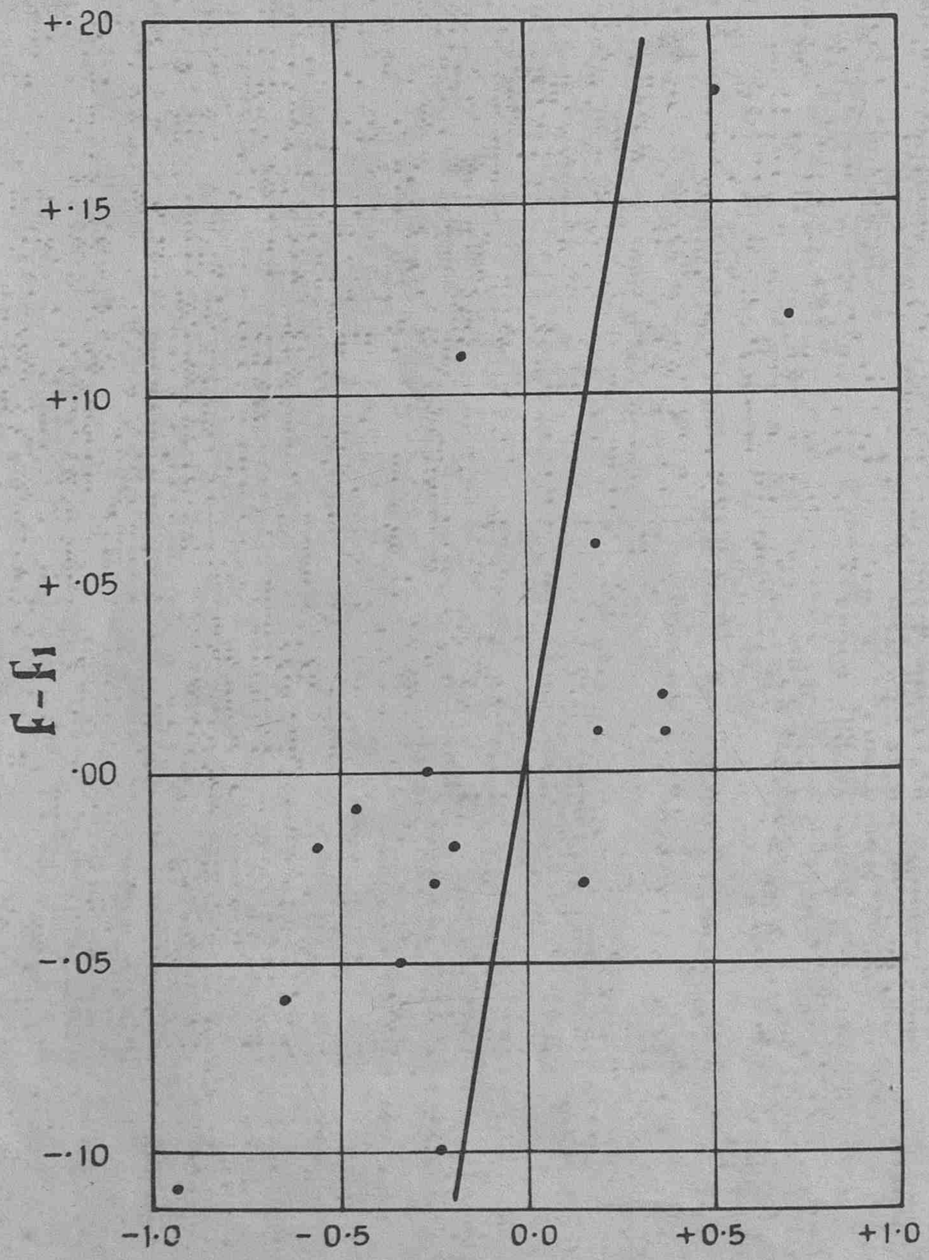

Fig. 2. Error at 16 Km.

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AIR MINISTRY MRP 204
20th October 1944
METEOROLOGICAL RESEARCH COMMITTEE

On the accuracy of Upper Air Observations

by

verre
S.Petterssen.

INTRODUCTORY NOTE.

This memorandum includes inter alia a comparison between the standards of performance of the U.S.A.A.F. and the Meteorological Office radio-sondes. In assessing the significance of the results obtained it should be borne in mind that the Meteorological Office instrument is used under more favourable conditions than is the U.S.A.A.F. instrument, i.e. the calibration of the Meteorological Office instrument is made at stations relatively near or at the radio-sonde station themselves, with the result that the time interval between calibration and use is relatively short.

Stornoway	7935	18009	29603
No. of obs. for each station.	172	170	129

4. The "wave length" of the pressure systems aloft and the rate of progress are such that the number of cases of reversal of sign of variation in height ought to be considerably smaller than the number of cases of persistency of sign.
The ratio of the number of reversal to the number of persistency is indicated in Table II.

/Table II

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AIR MINISTRY

MRP 204
20th October 1944

METEOROLOGICAL RESEARCH COMMITTEE

ON THE ACCURACY OF UPPER AIR OBSERVATIONS

(June 1944)

by

S.Petterssen

1. The network of upper air stations within the British Isles consists of Thums and Radio Sondes of (a) British, and (b) American make. Synoptic experience seems to indicate that the quality of observations is not uniform within the network, and the present note aims at assessing the discrepancies.

2. Comparison has been made between simultaneous observations from the stations indicated in the tables below. The observations referred to cover the period 27th March to 15th May, 1944 and each table represents a homogeneous series of observations.

3. Table I shows the mean heights of the 750, 500 and 300 mb. surfaces for the period under investigation. Plotting these heights on a chart and drawing the mean contours of the pressure surfaces, it will be seen that sets of smooth, or plausible, contours are not obtained. This shows that the inaccuracy of the observation is too large for a smoothing of the time series over 50 days to render smooth and plausibly spaced contours. On the other hand, it is not possible from such mean charts to indicate with certainty which stations are more reliable than others.

Table I.

Mean Heights (in feet) of Pressure Surfaces.

Station	750mbs.	500mbs.	300mbs.
Larkhill	8127	18325	30047
Liverpool	8062	18202	29862
Downham Market	8071	18216	29906
Lerwick	7878	17907	29702
Penzance	8133	18355	30074
Leuchars	7959	18099	29729
Stornoway	7935	18009	29603
No.of obs. for each station.	172	170	129

4. The "wave length" of the pressure systems aloft and the rate of progress are such that the number of cases of reversal of sign of variation in height ought to be considerably smaller than the number of cases of persistency of sign.
The ratio of the number of reversal to the number of persistency is indicated in Table II.

/Table II

Table II.

Percentage Number of Cases with Reversal of Sign to Number of Cases with Persistency of Sign, in the Height Variation of the 750, 500 and 300mb. Surfaces.

	%
Larkhill	48
Liverpool	58
Downham Market	53
Lerwick	63
Penzance	78
Leuchars	76
Stornoway	75

As far as the radio sonde stations are concerned it appears that the instruments of British make tend to render more plausible results than do those of American make. It will be seen that the difference between Penzance and Larkhill (both of which are in a fairly quiet regime) is appreciable. In individual cases, a difference in sign of the height variations at stations as near one another as are Larkhill and Penzance, tend to influence the conclusion that can be drawn from the analysed charts.

5. It may be assumed that there is no correlation between the errors of two consecutive observations. On this principle, it may be assumed that, other conditions being equal, a station that shows a small variability of heights is a better station than one that gives a large variability of heights.

Table III shows the mean 6-hour variability of heights at the various radio sonde stations.

Table III.

Mean 6-hour Variability in Heights of Pressure Surfaces (Feet)

	750	500	300mbs.
Larkhill	53	83	146
Lerwick	55	93	160
Downham Market	56	91	155
Liverpool	69	95	156
Penzance	64	117	211
Leuchars	67	117	214
Stornoway	75	127	221
No. of Obs. for each station.	172	170	129

Apart from the high variability at low levels at Liverpool (see para. 7.), it appears that the mean variability at British stations is considerably less than that at American stations. In particular, the large differences between such neighbouring stations as Larkhill and Penzance, and Stornoway and Lerwick indicate the difficulty that the analysts encounter in obtaining the contours with the accuracy required for determination of the winds aloft.

6. In relative terms, we may take Larkhill to be the "ideal" station and use its variability as a yardstick to measure the goodness of the other stations. Table IV shows the difference in variability between the other radio sonde stations and Larkhill.

/Table IV.

Table IV.

6-hour Variability minus the Larkhill 6-hour Variability(Feet)

	750	500	300
Larkhill	0	0	0
Lerwick	2	10	16
Downham Market	3	8	9
Liverpool	16	12	10
Penzance	13	34	65
Leuchars	14	34	68
Stornoway	22	44	75

It will be seen that the relative variability increases with elevation at all stations, except at Liverpool, and also that, on the whole, each of the two types of radio sondes give fairly uniform results. The high variaibility at American stations cannot easily be explained by reference to climatic or other conditions. Thus, there is nothing to indicate that Stornoway should have a 6-hour variaibility greatly in excess of that at Lerwick, nor can it be assumed that the circulation is such that Penzance should suffer much larger variations than Lerwick and Larkhill.

7. It is of interest to note that the variability at Liverpool is comparable with that at American stations at low levels and comparable with that at other British stations at high levels. The decrease along the vertical, as shown in Table IV, is most unusual and does not seem to suggest that the errors are mainly due to erroneous T-p relation, for inaccuracy in the T-p relation tends to give increasing height-error along the vertical.

Let V be the speed with which a pressure surface moves along the vertical, and b the corresponding pressure tendency.

Then

$$b = -\frac{\partial p}{\partial z} V$$

or

$$V = \frac{1}{\rho g} b$$

The mean variability of the height of a given pressure surface is then

$$\bar{V} = \frac{1}{\rho g} \bar{b}$$

since the variation in ρ in a given pressure surface is inappreciable.

The difference in the mean variability at Larkhill and Liverpool may be written

$$\Delta \bar{V} = \frac{1}{\rho g} \Delta \bar{b}$$

For two pressure surfaces, one above the other

$$\frac{\Delta \bar{V}_1}{\Delta \bar{V}_0} = \frac{\rho_0}{\rho_1} \frac{\Delta \bar{b}_1}{\Delta \bar{b}_0}$$

Table IV shows that the relative variability at Liverpool decreases with elevation, or

$$\frac{\Delta \bar{V}_1}{\Delta \bar{V}_0} < 1$$

Since

$$\rho_0 \geqslant \rho_1$$

it follows that $\Delta \bar{b}_0 \geqslant \Delta \bar{b}_1$

/It

METEOROLOGICAL RESEARCH COMMITTEE.

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The Errors of Upper Air Temperature and Pressure Measurements

by F.J. Scrase

(with an appendix by D.N. Harrison and A.W. Brewer).

1. Introduction.

The standard of accuracy for upper air temperature measurements, by radio-sonde and by aircraft ascents, now being aimed at by the Meteorological Office is ± ½°F at all heights up to at least 20 kms. The accuracy aimed at for pressure measurements is ± 2mbs down to 300mbs and ± 1mb at lower pressures. It was not expected that these stringent requirements could be obtained with existing instruments and new types of equipment are being developed which it is hoped will ultimately give the accuracy required. In the meantime efforts have been made to improve the performance of existing apparatus and technique and, what is more important, to obtain reliable assessments of the errors that occur in routine ascents. This latter subject has been approached from several different aspects and the object of this paper is to review and co-ordinate the results given in some recent reports and memoranda.

2. Sources of Instrumental Errors.

Measurements with the Kew (Mark I) radio-sonde are subject to the following errors:-

(a) Calibration errors. These are not likely to exceed 1.0°C at the lowest temperatures.

(b) Lag in response. For a normal lapse rate this causes an error of about +0.3°C near the ground during ascent but may increase to about 1°C at 200mbs.

(c) Radiation errors. These are discussed later.

(d) Receiver errors. These should be systematic at any one station.

(e) Hysteresis of pressure unit. This only enters to any appreciable extent during a descent since the instruments are calibrated with falling pressure.

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The instruments used for aircraft soundings are the aneroid barometer or altimeter (Mark XIV) and either a strut psychrometer or an electrical resistance thermometer. These instruments are subject to the following errors:-

(a) Calibration errors. These are controlled by specified tolerances.

(b) Errors in speed corrections to thermometer readings.

(c) Vibration errors. These are negligible except in the case of the earlier type (Mark I) of direct indicating electrical thermometer.

(d) Hysteresis of aneroid or altimeter and temperature compensation errors. Tolerances are specified and errors are small.

(e) Lag in response of thermometers. This is negligible in the case of electrical thermometers, and in the case of strut psychrometers is allowed for by level flying.

The tolerance in calibration error specified for the aneroid barometers is ± 2mbs and for the Mk. XIV altimeter ± 300 feet at about 500-300mbs. The instruments are checked at 3 monthly intervals to see that they are within the tolerances..

The strut thermometers are required to be correct to within ± 0.1°C down to 0°C and to within ± 0.2°C below that temperature. Tolerances for the electrical thermometers have not been laid down but the maximum accuracy is 0.5°C in the case of the earlier direct indicating pattern and 0.25°C in the case of the balanced bridge type.

The difficulty in assessing the overall performance of any particular type of instrument or method is to obtain a reliable standard against which the performance can be measured. The application of radar to height-finding is proving very useful in this respect.

3. Summary of Comparisons of Upper Air Measurement.

The comparisons which have been made during the past year can be divided into five main series as follows:-

(i) Differences of radio-sonde temperatures at various pressure levels, (a) on ascent and descent, (b) at midday and midnight, (c) at evening and early morning. These are discussed in M.R.P. 202 by E.G. Dymond.

(ii) Differences in aircraft temperatures, at various pressure levels, at midday and early morning. These are for Thum (strut psychrometer) ascents at Bircham Newton.

(iii) Comparison of radio-sonde and aircraft temperature at various pressure levels; (a) compares Downham Market R/S ascents with Bircham Newton Thum (strut psychrometer) ascents; (b) compares Larkhill radio-sonde ascents with Boscombe Down H.A.F. Spitfire (mainly electrical resistance thermometer) ascents.

(iv) Comparison of variability of heights of pressure levels indicated by radio-sonde and aircraft ascents. Full details are to be found in M.R.P. 204 by S. Petterssen.

(v) Comparison of heights computed from Larkhill radio-sonde observations with heights measured, on the same ascents, by G.L. III radar equipment. The full account is given in M.R.P. 203 by D.N. Harrison.

To facilitate discussion the main numerical results obtained in these comparisons are reproduced in a condensed form in Table I.

/Table I

Table I

Summary of Results of Comparisons.

Pressure level, mbs.		750	600	500	400	300	200	100	50
(i) Radio-sonde.		Temperature differences °C							
(a) Ascent-descent	(120)							-0.6	-0.3
(b) 1200-0000h.									
Summer: English Stations	(90)	+0.6	+0.8	+1.0	+1.2	+1.5	+2.0	+3.2	+5.7
Lerwick	(90)	+0.2	+0.3	+0.5	+0.6	+0.7	+1.1	+2.4	+5.0
Winter: English Stations	(90)	+0.3	+0.3	+0.3	+0.5	+0.9	+1.0	+1.7	+3.5
Lerwick	(90)	+0.3	+0.4	+0.5	+0.3	0	-0.2	+1.0	+3.1
(c) 1800-0600h.									
Autumn: Larkhill	(75)	+0.4	+0.4	+0.5	+0.3	0	-0.3	-1.0	+1.0
Winter: Larkhill	(75)	0	0	0	-0.1	+0.3	+0.3	+0.3	+0.3
(ii) Aircraft.									
(a) 1200-0600h. Bircham Newton Psychrometer:- Summer 1943	(60)	+0.3	+0.3	+0.2	+0.4				
Winter 1943-4	(60)	+0.1	-0.1	+0.1	0				
(b) Psychrometer - Electrical Therm. Aldergrove: Nov. 1943	(30)	0	+0.8	+1.0	+1.0				
(iii) Radio-sonde - Aircraft.									
(a) Downham - Bircham (Psych^{r}.) Oct. & Nov. 1943	(28)	+0.2	0	+0.3	+1.0				
(b) Larkhill - Boscombe Aug. 1943	(28)	+1.3	+1.1	+1.1	+1.8	+2.0			
April 1944	(22)	+0.7	+1.1	+1.0	+1.4	+1.8			
(c) Larkhill - Boscombe: Direct comparison	(2)	+0.5	+1.2	+1.5	+1.8				
(iv) Variability of Heights.		Height differences in metres.							
Aircraft	(70)	17		25					
Kew radio-sonde	(150)	17		28		47			
U.S. radio-sonde	(150)	21		37		65			
(v) Radio-sonde - Radar Heights. Larkhill 1944	(30)	-40	-90	-85	-70	-30	+20	-90	-100

Figures in brackets are the numbers of observations.

4. Radiation Errors in Radio-sonde observations.

Radiation errors can enter in two ways, (a) by the balloon becoming heated and leaving a heated wake of air in the path of the ascending radio-sonde and (b) by the radio sonde elements becoming heated by solar radiation. M.R.P. 202 shows by comparison of readings during ascent and descent (i(a) of Table I) that with the introduction of a 120 foot long suspension between the balloon and the radio-sonde the errors due to (a), which had previously been as much as +3°C, (M.R.P.67) have now been eliminated. The small negative difference of about -0.5°C, now found between ascent and descent is probably due to battery failure on a high ascent.

It is not quite so certain, however, that improved radiation shields have entirely eliminated errors due to (b). Differences in midday and midnight temperatures recorded at the English radio-sonde stations in summer (i(b) of Table I) of up to 2°C in the troposphere and increasing to about 6°C. in the stratosphere, are little less than those found before additional shielding was fitted (M.R.P.67). Comparisons, however, of the results obtained at 0600 and 1800 hrs., when solar elevation is about the same and radiation errors should therefore be the same, give some indications of a true diurnal variation of about 0.5°C. between these two times of day, but further observations are needed in order to eliminate the effect of change of air mass. It is of interest to note here that W.H.Dines (Q.J.R.Met. Soc., 40, 1944, 1-12) compared temperatures recorded by balloon meteorographs at sunrise and sunset and found a difference of about +1°C. (sunset-sunrise) from 2 km. up to 8 km.; the similarity to the recent result is noteworthy.

On the other hand, differences between 0600 and 1200 hr. aircraft observations at Bircham Newton for summer and winter (ii(a) of Table I) are considerably smaller than the balloon observations. For comparison the latter are plotted with these aircraft observations in Fig.I(d). The difference may be due partly to the shorter time interval between the observations (midnight soundings are not made at Bircham Newton) and partly to a smaller radiation effect. The much higher ventilation speeds of aircraft thermometers make radiation effects much less important.

The difference between the English stations and Lerwick, illustrated in M.R.P. 202 and in Fig 1., suggests either that solar elevation at midday is an important factor or that if there is a true diurnal variation of temperature aloft it is affected by the extent of land mass or sea below.

On the whole the evidence provided by these comparisons suggests that there is a real diurnal variation of temperature in summer up to about 400 mbs. but not in winter above 800 mbs. The existence of a diurnal variation of heights above about 350 mb. level is supported by the behaviour of decrease in the amount of very high cloud, in the afternoon (M.R.P.36), by condensation trail data for day and night, by the comparison of trails at Wick with those at Boscombe Down and by the fact that persistent trails in the stratosphere have been observed almost solely in the early morning (M.R.P.99).

Before leaving this question of diurnal variation it should be pointed out that the differences of temperature under consideration are those corresponding to given pressure levels. If variations at the same geometric height are considered then, quite apart from any diurnal variation in temperature due to other causes there will be a variation due to any diurnal variation of pressure at the given height. A contraction of the atmosphere below the given geometric height will result in air from above that height falling without appreciable change in its pressure and therefore retaining the temperature from the higher level.

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5. Aircraft and Radio-sonde Comparisons.

Some comparisons of observations made in Spitfire aircraft, using electrical and other thermometers, at Boscombe Down with radio-sonde observations at Larkhill in August 1943 (Table I, iii(b)) showed consistent positive differences of between 1° and 2°C, the radio-sonde giving higher readings. This was confirmed by a further series of comparisons in April 1944 after there had been an improvement in the procedure for making the aircraft observations, including the use of the balanced-bridge electrical thermometer. On the other hand, comparisons of aircraft observations, using strut psychrometers, at Bircham Newton with radio-sonde observations at Downham Market showed practically no difference up to 500 mbs. but some moderate departures occurred at lower pressures. All these comparisons are illustrated in Fig. I(a) and (b). From these results it might be inferred that the radio-sonde and strut psychrometer temperatures are more reliable than the electrical thermometer temperatures. That there is a tendency for the earlier pattern of electrical aircraft thermometer to give lower readings than the psychrometer is borne out by a comparison of aircraft observations (Thum and Prata) with both types at Aldergrove in November 1943 (see ii(b) of Table I). There is good reason to believe that the balanced-bridge type of electrical thermometer gives closer agreement.

All these results are based on statistical comparisons of observations not necessarily having very close agreement in time and place and this might well account for some of the discrepancies. Recently two direct comparisons have been made by an aircraft circling round an ascending radio-sonde, thus eliminating time and place differences. A brief report on these ascents, by D.N.Harrison and A.W.Brewer, is reproduced as an Appendix to this paper. Owing to the limited rate of climb of the aircraft, the rate of ascent had to be reduced from the normal of 1200 ft./min. to 890 ft./min in the first ascent and 345 ft./min in the second. The results are plotted in Fig. I(c). A balanced-bridge thermometer was used for the aircraft observations.

In the first ascent there were no systematic or large differences up to about 600 mbs. but an increasing positive difference (radio-sonde readings higher) above that level. These results, in fact, are practically the same as the Downham Market-Bircham Newton results referred to above. In the second ascent the radio-sonde temperatures were systematically high by about 1°C. up to 600 mbs. and 2°C. at higher levels. There was no systematic difference between the pressure observations, so there is no question of the temperature differences being due to pressure errors. It seems fairly certain that they must be due to radiation error which one would expect to be more pronounced with the low rate of ascent. They are not likely to be due to the increase in the lag coefficient of the thermometer element with decrease in rate of ascent since this is more than counteracted by the decrease in the rate of change of temperature with time, the actual lag in temperature, $\theta_o - \theta_c$, being roughly:-

$$\theta_o - \theta_c = -\frac{\partial\theta}{\partial h} V \lambda$$

$\partial\theta/\partial h$ being the lapse rate, V the rate of ascent and λ the lag coefficient which is proportional to V^n where n is probably about 0.7.

If one can draw a general conclusion from the radio-sonde aircraft comparisons as a whole it is that the temperature differences up to about 500 mbs. are roughly as follows:-

	°C.
Radio-sonde - electrical thermometer	+0.7
Radio-sonde - psychrometer	0
Electrical thermometer - psychrometer	+0.7

At higher levels there is a general tendency for the radio-sonde to give increasingly higher readings and this is most likely due to radiation error. The differences at the lower levels might be due to errors in the speed corrections to the aircraft psychrometer readings or to differences in lag in the three cases, the lag of the electrical thermometer being less than the lag of the psychrometer and the radio-sonde.

6. Variability of Upper Air Observations.

In M.R.P. 204 S. Petterssen has assessed the relative reliability of the observations at the various stations in the British Isles by comparing the variability, over 6 hour intervals, of heights of the 750, 500 and 300 mb. pressure surfaces. On this basis the aircraft (psychrometer) observations are slightly more accurate than the British radio-sonde observations and the latter more accurate than the observations with U.S. radio-sondes. It should be remembered that the latter are not individually calibrated, as are the Kew instruments. As is to be expected the variabilities increase with height. In one case, Liverpool, the value for 750 mbs. is high compared with the other British stations. It is understood, however, that this anomaly, which might have been due to non-systematic errors in surface pressure, has disappeared in a more recent series of observations.

As the variability is due partly to the true changes in the heights of the pressure surfaces and partly to random errors it is of interest to try to assess the two parts. If Δh_1 is the average error of a single determination in the height the average error of the difference between consecutive observations is $\sqrt{2}\,\Delta h_1$. Then if Δh_c is the mean true change in height over a 6 hour interval (i.e. the true variability), the mean apparent variability ΔH is given by

$$\Delta H^2 = \Delta h_c^2 + 2\Delta h_1^2$$

$$\text{But} \quad \Delta h_c = \Delta p_c \frac{\partial h}{\partial p} = \frac{\Delta p_c}{\rho g} = \frac{RT}{gp}\Delta p_c$$

where Δp_c is the pressure change corresponding to Δh_c, ρ the air density and T the absolute temperature at pressure level p; g is gravity and R is the gas constant. may also write:-

$$\Delta p_c = \Delta p_o \frac{p}{p_o} \qquad \text{approximately,}$$

where Δp_o is the mean pressure change (in 6 hours) at ground level pressure p_o.

$$\text{Therefore} \quad \Delta h_c = \frac{RT\,\Delta p_o}{g\,p_o}$$

For Δp_o we may take 1.8 mbs. the average change of surface pressure in a 6 hour interval at Croydon in the first half of 1944 (computed by Dr. Brooks). This gives Δh_c (in metres) $= 0.52$ T.

Using this and average values of T with the average variabilities of the British radio-sonde observations we obtain the following results:-

Table II.

Pressure Level mbs.	Apparent Variability ΔH m.	T °A.	True Variability Δh_c m.	Mean error Δh_1 m.
750	17	272	14	7
500	28	252	13	17
300	47	227	12	32

Thus the mean random errors in heights increase from about 25 feet at 750 mbs. to about 100 feet at 300 mbs.

7. Comparisons of Heights measured by Radar and Radio-sonde.

The most reliable check of radio-sonde accuracy that has yet been made is that described by D.N. Harrison in M.R.P. 203 in which heights derived from Kew Mark I radio-sonde observations of pressure and temperature are compared with those measured by radar (GL.III) observations of slant range and elevation of a reflector carried by the same balloon. Comparison of the radar method against visual theodolite observations, which, after taking account of scale errors, were found to have a probable error of about 0.01° in azimuth and elevation, showed that the probable error of a single reading in elevation on the radar set was less than 0.1°. The range accuracy is of the order of 10 metres and under average conditions of range and elevation the probable error of a single observation of height works out at about 30 metres. In Harrison's comparison such errors have been largely removed by smoothing in computation and it is concluded that the radar heights are sufficiently accurate to be used for estimating the errors of the radio-sonde heights.

The main feature of Harrison's results, some of which are reproduced in graphical form in Fig. 2, is a systematic negative error of up to 90 m. in the radio-sonde heights in the first 9km. There is good agreement between the results obtained in 1943 and those obtained in 1944. The error appears to be partly due to imperfect temperature corrections to the readings of the pressure unit since there is some slight evidence, in Figs. 1 and 2. of M.R.P.203, of correlation between the magnitude of the error and the correction factor. This, however, does not explain the whole of the height error.

The error becomes positive at about 10 km. with a sharp maximum of +0.02 km. at the tropopause. This suggests a lag of the temperature of the pressure unit behind that of the temperature unit (it is the temperature indicated by the latter that is used in obtaining the pressure unit corrections). The effect, in a positive lapse rate, is to give negative pressure errors which are greater for a greater temperature lag of the inductances. There is no evidence of systematic error in the stratosphere, which again suggests that the error in the troposphere is associated with temperature lag of the pressure unit. Against this, however, is the fact that measurements of the temperature effects and lag coefficients of the various parts of the unit do not indicate any perceptible source of error.

Harrison has considered in some detail the effect of temperature errors on the radio-sonde heights. It is complicated by the fact that by day, even if the effect of differences in lag of the temperature and pressure units is inappreciable, different parts of the units will be affected unequally by radiation. Comparisons of radio-sonde and radar height differences by day and by night may throw some light on this question.

In considering the difference which Harrison found between the measurements by the two methods it is important to remember that they refer to heights at given times and not to heights of given pressure levels, the errors of which should, in general, be considerably smaller.

The error in the height of a given pressure level computed from a pressure-temperature sounding is

$$\Delta h, = \frac{R}{g} \left(\overline{\Delta T} - \overline{\frac{\partial T}{\partial p}\Delta p} \right) \log p_0/p \qquad \text{--------(i)}$$

Where $\overline{\Delta T}$ is the error in the mean temperature of the layer between the pressures p_0 and p and $\overline{\frac{\partial t}{\partial p}\Delta p}$ is the mean equivalent error in temperature due to an error, Δp, in pressure. On the other hand, when the height at a given time, or at a significant point on the p - T curve such as the tropopause, is computed, an additional, and normally larger error due to difference between the indicated pressure and the true pressure at the given time, is involved. This additional error is $\frac{\partial h}{\partial p}\Delta p$ and the total error is then

$$\Delta h_2 = \Delta h_1 + \frac{\partial h}{\partial p}\Delta p \quad \ldots\ldots\ldots(2)$$

or since $\frac{\partial h}{\partial p}\Delta p = -\frac{RT}{gp}\Delta p$

$$\Delta h_2 = \frac{R}{g}\left[\left(\overline{\Delta T} - \frac{\partial T}{\partial p}\Delta p\right)\log\frac{p_0}{p} - \frac{T}{p}\Delta p\right] \quad \ldots\ldots\ldots(3)$$

The last term of this expression is normally much larger than the other terms and so Δh_2 is usually much greater than Δh_1. Thus the mean random errors given in columns 5 and 6 of Tables I and II in M.R.P. 203 compare with the mean errors derived for the pressure levels in Table II of Section 6 as follows,

Table III.

Pressure level mbs.	Mean error Δh_1 m.	Height km.	Mean error Δh_2 m.	Mean error Δp mbs.
750	7	2.5	70	6.5
500	17	5.5	90	6.0
300	32	9	130	5.5

The mean pressure errors given in the last column are derived from the difference between Δh_2 and Δh_1 (squares being used since mean random errors are being considered). Since the effect of Δh_1 on Δh_2 is small they are practically the same as the values given in the last column of Table II of M.R.P. 203 which were computed by assuming Δh_1 was negligible. These random pressure errors are rather greater than was previously assessed as ± 4 mbs. probable error by comparing two pressure elements on the same transmitter but since the latter comparison eliminates any temperature errors the agreement is reasonably satisfactory.

It should be noted that above the tropopause, where $\partial T/\partial p$ is negligibly small, the height errors of given pressure levels are almost entirely due to temperature errors and the large errors computed in the fourth column of Table VIII of M.R.P. 203 would not, in fact, be evident if pressure levels rather than significant points were being considered.

Although the radio-sonde-radar comparisons do not, by themselves, enable us to separate the errors of temperature from the errors of pressure it is possible to obtain a little information about the former by using the derived values of Δp and Δh_1 together with average values of $\partial T/\partial p$ appropriate to the pressure levels, to obtain values of $\overline{\Delta T}$ from the expression (1). The following result is obtained:-

Table IV.

Pressure level mbs.	$\Delta h_1 / \frac{R}{g}\log\frac{p_0}{p}$ °C	$\frac{\partial T}{\partial p}$ °C/mb.	$\frac{\partial T}{\partial p}\Delta p$ °C	$\overline{\Delta T}$ °C
750	.85	.05	.3	.8
500	.85	.10	.5	.7
300	.9	.13	.55	.7

Thus the random error in mean temperatures up to the levels concerned is about 0.75°C; the random error in observed temperatures at these levels is therefore a little over 1°C. This is in fairly satisfactory agreement with a result obtained from readings of two thermometer units on the same transmitter, which showed agreement to within about 0.8°C. It appears, therefore, that the deductions made about the random instrumental errors from the variability results are consistent with the results from the radar comparisons.

Summary.

The Kew (Mark I.) radio-sonde is not entirely free from radiation error. This may amount to about 1.5°C. in the troposphere and more in the stratosphere. There is evidence of a real diurnal variation of temperature up to at least 400 mbs. in summer (probably not exceeding 1°C. at that height) and up to perhaps 800 mbs. in winter; the differences between day and night observations in the higher troposphere and in the stratosphere require further investigation.

Radio-sonde temperatures and aircraft psychrometer temperatures are about 0.7°C. higher than those obtained with aircraft electrical thermometers.

Random mean errors in height of pressure levels increase from about 25 feet at 750 mbs. to about 100 feet at 300 mbs. Random mean errors in radio-sonde pressures and temperatures do not exceed 6mbs. and 1°C. respectively.

Systematic errors of up to 90 m.. in height derived from radio-sonde observations are partly due to imperfect temperature corrections to the pressure readings. No systematic errors have been observed in the stratosphere.

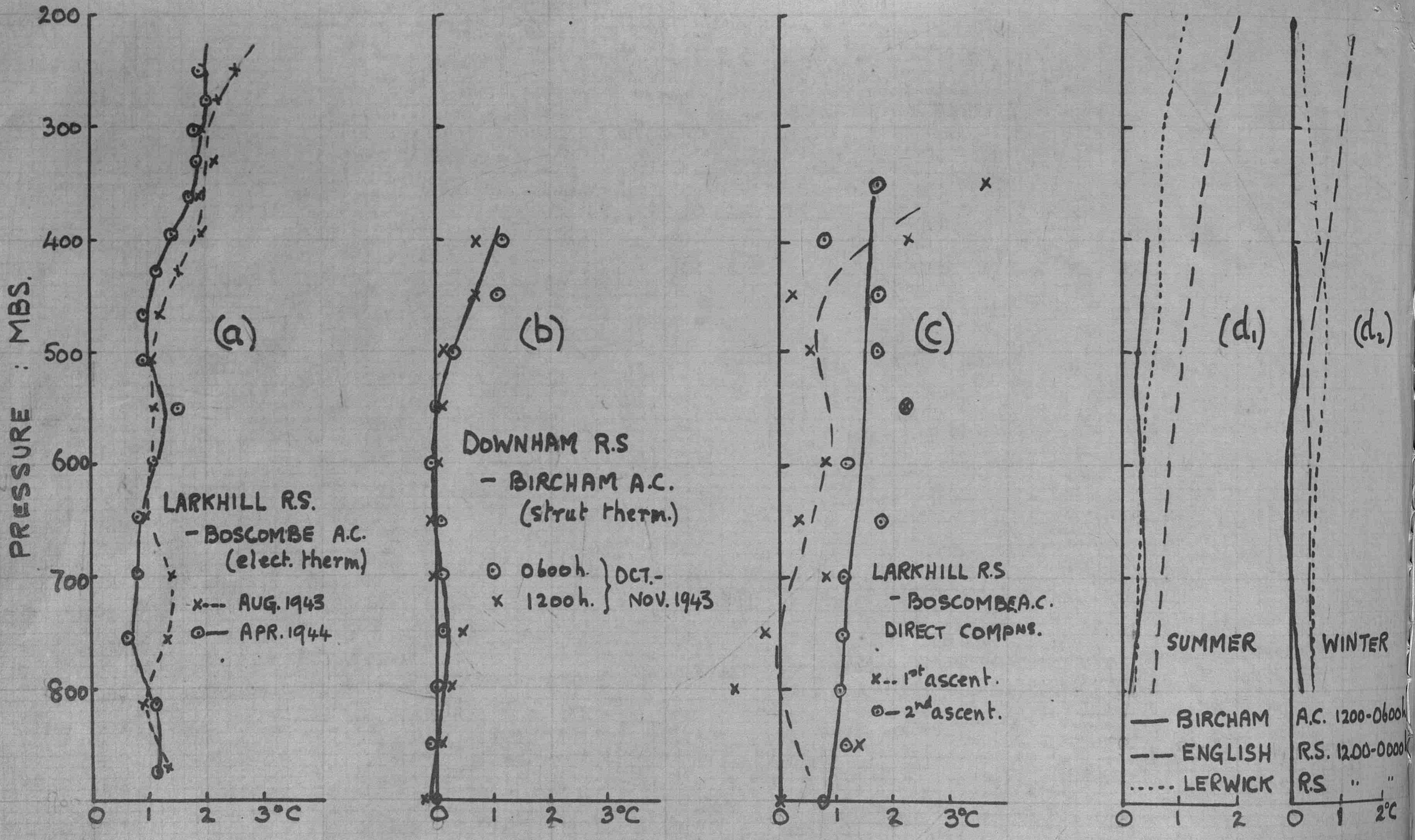
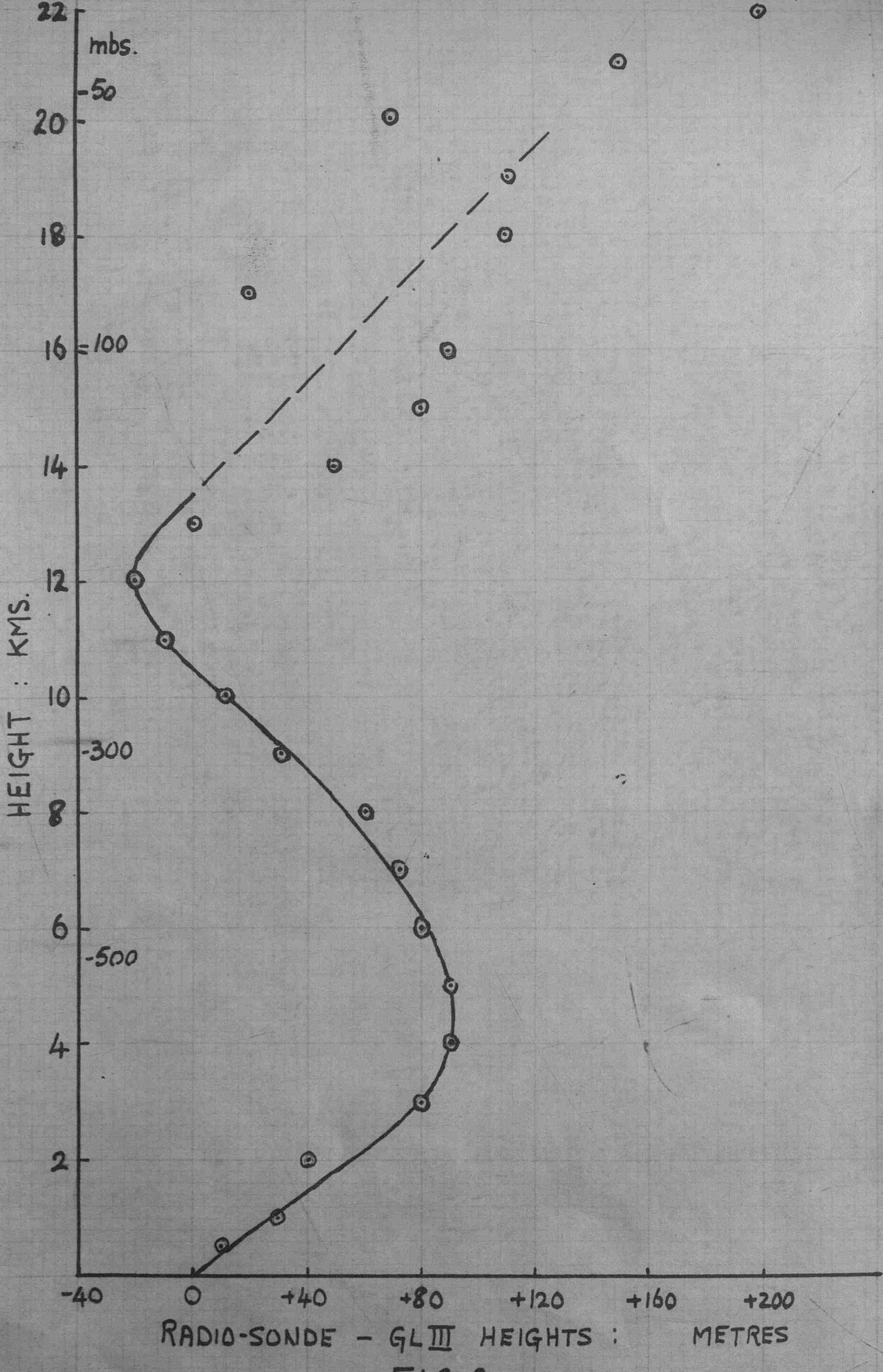


FIG. 1. RADIO-SONDE AND AIRCRAFT COMPARISONS.

MRP 205



22
mbs.
-50
20
18
16
=100
14
12
10
-300
8
6
-500
4
2
HEIGHT : KMS.
-40
0
+40
+80
+120
+160
+200
RADIO-SONDE - GL III HEIGHTS : METRES

FIG. 2.

SECRET.

Appendix I - A Direct Comparison of Aircraft and Radio Sonde Temperature and Humidity measurements.

D.N. Harrison
A.W. Brewer.

~~SECRET~~

Introduction.

Recently there have been very considerable improvements in upper air temperatures and humidity measurement from aircraft and it has been felt that a direct comparison between a Radio Sonde and an aircraft circling the balloon would be of value in checking the accuracy which is being obtained by Radio Sonde methods. The balloon is circled by the accompanying aircraft to eliminate the effect of differences in place and time of two independent soundings.

This report presents the results which have been made on two such ascents.

The Aircraft observations.

The aircraft used was the Fortress allotted to H.A.F. for Meteorological work. Pressure measurements were by Mk.XIV altimeter recently calibrated and with no corrections or hysteresis exceeding 1 mb. A height/G.M.T. log was kept on the second ascent to compare with the Radio Sonde pressure/G.M.T. record. The temperature measurements were made on a balanced bridge electrical resistance thermometer, which is a simple and exceedingly stable instrument free from any significant errors due to voltage or ambient temperature changes, using a thermometer element of which when ventilated at flying speeds the lag is too small to measure by ordinary methods. Speed corrections have to be applied which vary from 4 - 10°F. The error in the speed correction should not exceed 10% and is usually much less.

The humidity measurements were by Dobson frost point hygrometer in which the frost point is measured directly and independent of speed corrections. Laboratory experiments have shown that it is practicable to measure a humidity at -80°C to an accuracy of about 10%. This accuracy is not attainable in the air but the loss for this purpose is not significant; it may be noted that on account of the principle of this hygrometer it is possible at a temperature of +30°F to discriminate between 0.9% and 1.0% humidity, as this corresponds to a change of 3°F in the frost point.

The Radio Sonde observations.

These were made in the normal way, except that (a) the rate of ascent was reduced to that at which the aircraft could climb and (b) the number of readings taken was greater than usual. A large reduction in the rate of ascent is usually considered objectionable, but some reduction was inevitable. At the first trial the rate was only slightly below normal, but for the second it was reduced still further. In order to show the effect of this low rate of ascent, the temperatures at the standard pressure levels, together with those recorded at the preceding (06h) and following (12h) routine ascent, are given in Table I. In nearly every case the temperature at this (09h) ascent lies between the 06h and the 12h value. It appears therefore that the effect of insolation must be small, but it is perhaps worth noting that the 06h temperatures are almost exactly the same as those measured by the aircraft at 09h.

The points measured on the R-S record are those where dT/dp or dU/dp show marked changes. These are the values given in the tables. Intermediate values may be found by linear interpolation.

Table I

Radio-sonde temperatures for normal and low rate of ascent.

Time of ascent G.M.T.	06	09	12
Mean rate of ascent (ft/min)	1330	345	1130
P (mb)	T(°F)		
950	49	49	47
900	51	50	49
850	47	49	45
800	41	43	44
750	34	36	38
700	25	29	30
650	20	23	25
600	13	17	19
550	7	10	12
500	-1	+3	+3
450	-12	-5	-4
400	-24	-20	-18
350	-36	-32	-30
Mean	16.5	19.4	20.0

Ascent No.1. 28 April 1944. 09.27 G.M.T.
Weather: cy, Sc 4/10 + Ci
Mean rate of ascent to 400 mb, 890 ft/min.

The high rate of ascent, combined with inexperience of this type of ascent, resulted in no time - height record being kept. Owing to the reduced time available for humidity measurements there was a slight loss of accuracy, but it is considered that this was insignificant in comparison with the differences between the aircraft and radio-sonde observations.

In the R-S observations there was a jump of about +6 cycles/sec in all elements at 435 mb. The resulting uncertainty in the subsequent measurements is about +3 mb in P, +1°F in T and +1% in U.

The results of this ascent are given in Table II and Fig.1.

The temperatures are in very good agreement up to 650 mb, after which the R-S values are the higher, the difference reaching 5½°F at 375 mb.

In the case of humidity there are large differences in the lower levels. The gross discrepancy between aircraft and radio-sonde at 900 mb is presumably due to aircraft's selection of clear air.

At 595 mb the aircraft recorded a dry layer between two moist layers in which, on account of a heavy ice deposit, the actual humidity could not be measured. The radio-sonde hygrometer shows a gradual increase of U between 670 mb and 570 mb, but does not reproduce the rapid variations shown by the aircraft readings. Above this level also the R-S record is smoother than that of the aircraft. This suggests the effect of lag.

There is a marked decrease in the rate of ascent of the balloon at 530 mb, for which no reason can be assigned.

Ascent No.2. 1 May 1944, 09.20 G.M.T.
Weather: bcm$_o$, Fs 7/10 at 1200 ft.
Mean rate of ascent to 400 mb, 345 ft/min.

The low rate of ascent reacts favourably on the aircraft observations, which are all satisfactory.

The results of this ascent are given in Tables III and IV and Fig. 2. The R-S temperatures are all higher than the aircraft's after 940 mb; the difference is about 2°F between 900 mb and 600 mb, and 3° to 4°F after 600 mb. The apparent temperature differences cannot be ascribed to pressure differences.

The humidity changes were less violent than at the previous ascent, and the lowest humidity recorded is 7%. The radio-sonde shows a curve similar to the aircraft, but the changes are smoothed, particularly at high levels, and throughout the ascent the R-S values are considerably the higher.

The differences between the aircraft pressure readings and the R-S values for the same times are given in Table IV. The R-S pressures are interpolations, read off the pressure-time curve of Fig. 2, since the aircraft readings could not be made to coincide with the pressure signals from the transmitter. In the construction of the pressure-time curve the values given in Table IIIhave been supplemented by others computed as far as possible from the actual frequency readings.

Future work. At present no aircraft is available to continue this work.

Table II - T and U comparison.

ASCENT No.I. 28.4.44.
09.27 G.M.T.

AIRCRAFT						RADIO-SONDE				
Ht. (ICAN)	P	T	Ts	U	Remarks	Time	P	T	P	U
ft.	mb	°F	°F	%		mins	mb	°F	mb	%
1000	973	42	34	77		0.0	1013	51	-	61
3000	908	33	-15	9	Poor U	1.6	934	36½	898	76
5000	843	36	-28	4		2.7	890	31½	856	30
6000	812	38	-54	1		3.7	856	39	747	15
7000	782	34	-53	1		5.1	824	36	725	14
8000	753	31	-33	3		5.8	806	36	678	12
9000	724	29	-55	1		8.2	740	29½	673	10
11000	670	28	-9	19		8.8	723	30	655	17
13000	620	20	-		Humidity high-unknown	10.7	666	27½	605	26
14000	595	16	-16	19		13.7	582	16	582	48
15000	572	9			Humidity high-unknown	15.3	538	10½	563	51
17000	527	7	-9	43		17.3	507	6	496	53
18000	506	4	-11	45		18.7	492	1½	473	51
19000	485	-2	-15	49		20.1	475	-1½	461	55
20000	465	-5	-20	43		21.1	462	-5½	420	54
21000	446	-10	-21	54		25.4	411	-15	391	48
22000	428	-14	-27	47		30.7	354	-28½	376	48
23000	409	-20	-35	40		31.3	349	-29½	347	41
24000	392	-24	-36	48		31.9	343	-32		
25000	375	-29	-36	65	Good measurement taken at leisure					

/Table III

Table III - T and U comparison

ASCENT No.2. 1.5.44.
09.20 G.M.T.

AIRCRAFT						RADIO-SONDE				
Ht. (ICAN)	P	T	Ts	U	Remarks	Time	P	T	P	U
ft	mb	°F	°F	%		mins	mb	°F	mb	%
1000	977	48	43	17		0.0	1008	55	-	77
2000	942	52	52	100	In & out of cloud	2.6	975	47	971	84
3000	908	50	30	12		3.4	960	45½	952	81
4000	875	48	6	22		4.8	940	52	938	69
5000	843	47	6	15	Very variable, just entered cloud or haze layer.	9.6	884	50	912	65
6000	812	42	21	16		12.9	844	48½	902	61
7000	782	39	11	41		16.5	812	43½	878	42
8000	752	34	6	29	Very variable	21.4	762	38½	868	39
9000	724	30	5	26	Good	23.6	741	34	840	38
10000	696	26.5	4	30	"	27.5	714	31½	803	54
11000	670	23	0	34	"	31.5	675	25	751	44
12000	644	19.5	-7	32	"	39.0	613	20	688	48
13000	619	18	-23	26	"	43.6	569	10	651	48
14000	595	14	-4	17	Variable	48.2	538	9½	604	41
15000	571	9	-4	40		48.9	534	8	596	45
16000	549	6	-17	29		54.0	495	2	562	49
17000	527	3	-15	38	Good	56.7	469	-2	525	43
18000	505	1	-33	14	"	61.5	441	-7½	454	33
19000	485	-3.5	-32	19	"	63.7	422	-15	431	31
20000	465	-6	-52	7	"	67.4	394	-21½	411	31
21000	446	-10	-54	7	"	71.0	371	-26	358	35
22000	427	-15	-42	21	"	78.5	325	-39	350	34
23000	404	-21	-46	14	Moderate				549	44
24000	392	-25	-39	40						
25000	375	-29	-43	40						

Table IV - P comparison.

ASCENT No.2.

Time	P (mb)	
mins	Aircraft.	RS-aircraft.
2.3	977	+2
6.05	929	-4
10.3	872	+2
20.05	776	0
22.2	752	+3
34.8	644	-12
45.05	549	+10
53.25	511	-10
60.2	446	+5
69.25	375	+7

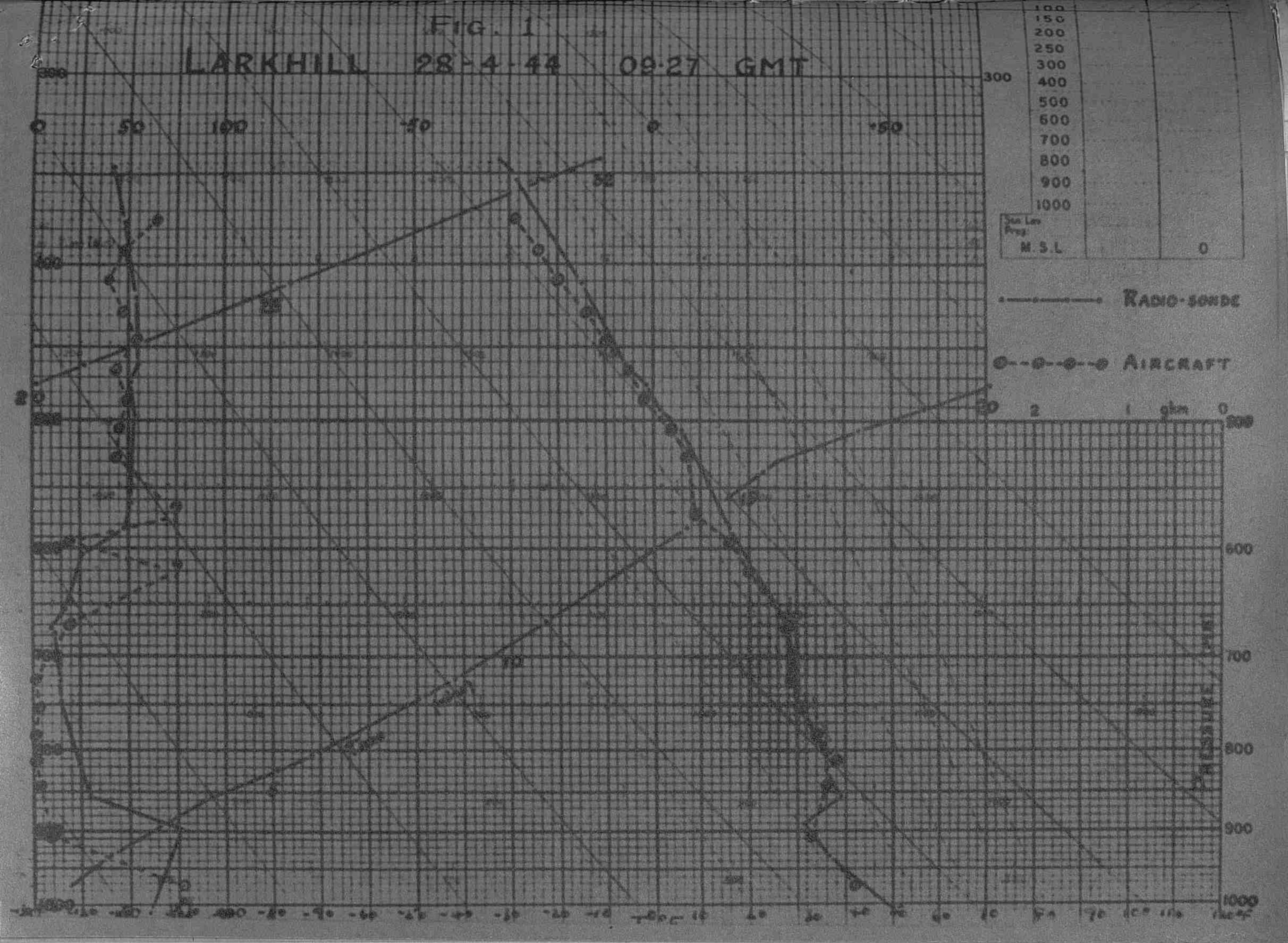


FIG. 1

LARKHILL 28-4-44 09·27 GMT

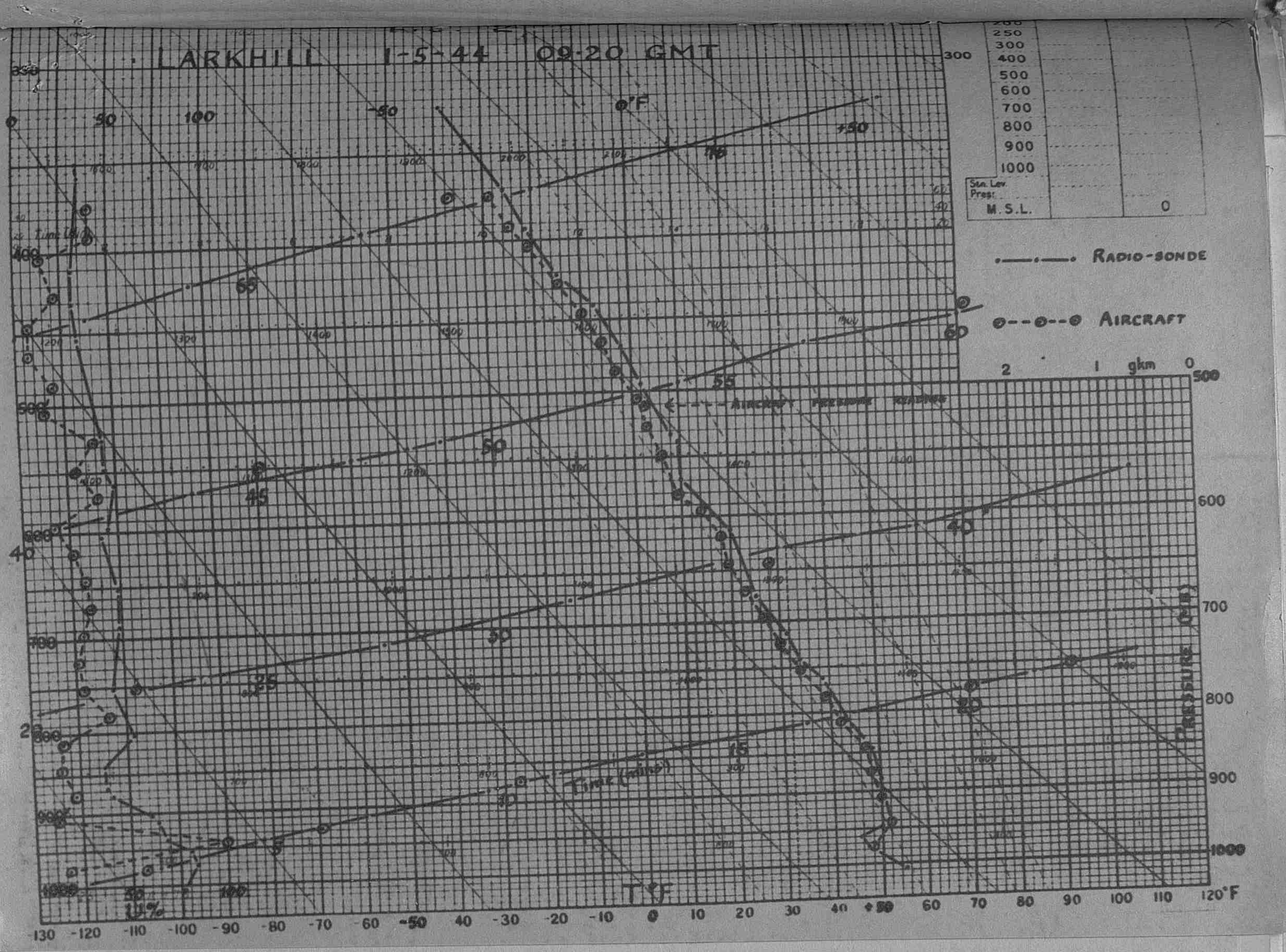
Date	12	14	19	20	21	23 July 1943
Hygro"T"	0	+11.5	-2	+5.5	+5	+8.5 c/s
Hygro"U"	+6	+11.5	-14	+2.0	-8.5	-7.0

On the 14th, corrections on one element were nearly double the extreme values allowed in routine work. The elements were used however in this condition to test the effect of large corrections. The largest deviation of the two elements in the ascent was 14% at 300 mbs. If account is taken of the systematic difference between T and U position (see below) this reduces to 5%. In this instance therefore the control correction although large has yielded good results. But it cannot be assumed that it will do so in all cases, as the effectiveness of the corrections must depend on a proper knowledge of the shape of the calibration curve.

Results of Ascents.

These are shown in the accompanying figures, where in all cases the dotted trace is that of the element in the normal U position, the full line that in the temperature T position.

A casual inspection shows that there is a systematic difference between the two curves. This is not unexpected. The U position in the instrument is known to suffer from radiation, which may in both cases [illegible]



Date	12	14	19	20	21	22	[illegible]
Hygro"M"	0	+12.5	-2	+5.5	+5	+6.5	c/s
Hygro"U"	+6	+18.5	-4	+2.0	-5.5	-7.0	

On the 14th, corrections on one element were nearly double the extreme values allowed in routine work. The elements were used however in this condition to test the effect of large corrections. The largest deviation of the two elements in the ascent was 24% at 500 mbs. If account is taken of the systematic difference between M and U position (see below) this reduces to [illegible]%. In this instance [illegible] the control correction although large [illegible] adequate. But it cannot be assumed that it will be so in all cases, as the [illegible] of the corrections [illegible] large changes in the shape of the calibration curve.

[illegible]

[illegible] In all cases the [illegible] trace is that of the element in the normal U position, the [illegible] line that in the experimental M position.

A careful inspection shows that there is a systematic difference between the two curves. This is not unexpected. The U position in the [illegible] is known to suffer from [illegible]

METEOROLOGICAL RESEARCH COMMITTEE

Humidity Measurements with the Kew Radiosonde.

by

E.G.Dymond.

The accuracy of the Kew Radiosonde has been checked in the past by series of ascents of special instruments with either two barometers, or two thermometers. This report describes the result of a similar series of six ascents with two hygrometers.

Transmitters were given normal routine calibrations with one hygro element in the T (thermometer) and one in the standard U position. The two elements of any one transmitter were not simultaneously calibrated in one vessel, so normal calibration errors are included in the results.

Calibrations took place from one to forty-two days before ascent, though the majority were about a week old. No correlation was found between age of calibration and consistency of readings, within the limits experienced.

As no temperature element was fitted, the corrections to the pressure element were computed from the temperatures of the Larkhill 1200 ascent, which was always within three hours of the experimental soundings.

Control corrections.

These are given in Table I. Roughly speaking, 1c/s is equivalent to 1 % R.H.

Table I.

Date	12	14	19	20	21	23	July 1943
Hygro"T"	0	+18.5	-2	+5.5	+3	+8.5	c/s
Hygro"U"	+6	+38.5	-14	+2.0	-6.5	-7.0	

On the 14th, corrections on one element were nearly double the extreme values allowed in routine work. The elements were used however in this condition to test the effect of large corrections. The largest deviation of the two elements in the ascent was 14% at 300 mbs. If account is taken of the systematic difference between T and U position (see below) this reduces to 5% . In this instance therefore the control correction although large has yielded good results. But it cannot be assumed that it will do so in all cases, as the effectiveness of the corrections must depend to a large extent on the shape of the calibration curve.

Results of ascents.

These are shown in the accompanying figures, where in all cases the dotted trace is that of the element in the normal U position, the full line that in the temperature T position.

A casual inspection shows that there is a systematic difference between the two curves. This is not unexpected. The U position in the transmitter is known to suffer from interaction, which may in bad cases change the frequency, between calibration and flight, by as much as ten cycles. The T position does not suffer from this. The full line curves in the figures are therefore more likely to be free of systematic errors than the dotted curves. This systematic error has since been removed and will therefore not occur in the humidity readings obtained in a normal ascent.

In the assessment of the consistency, or casual errors, of the hygrometer there is justification for eliminating this systematic difference, which increases with increasing height. It is not directly a function of height but generally of departure from control conditions.

The systematic differences were evaluated by taking mean values at different levels paying attention to the sign.

The average difference between two elements, for all ascents is shewn in Table II.

Table II.

Level.	A.	B.	C.
900 mbs.	4%	3	10
800	6	3	8
700	4	3	8
600	6	4	9
500	7	4	7
400	7	4	9
300	9	6	9
200	9	4	9

Column A gives mean difference in percent ratio of humidity regardless of sign, of the observed values.

Column B gives mean difference also without respect to sign when correction has been made for the systematic difference.

Column C gives the maximum difference in any ascent, after correction for the systematic effect.

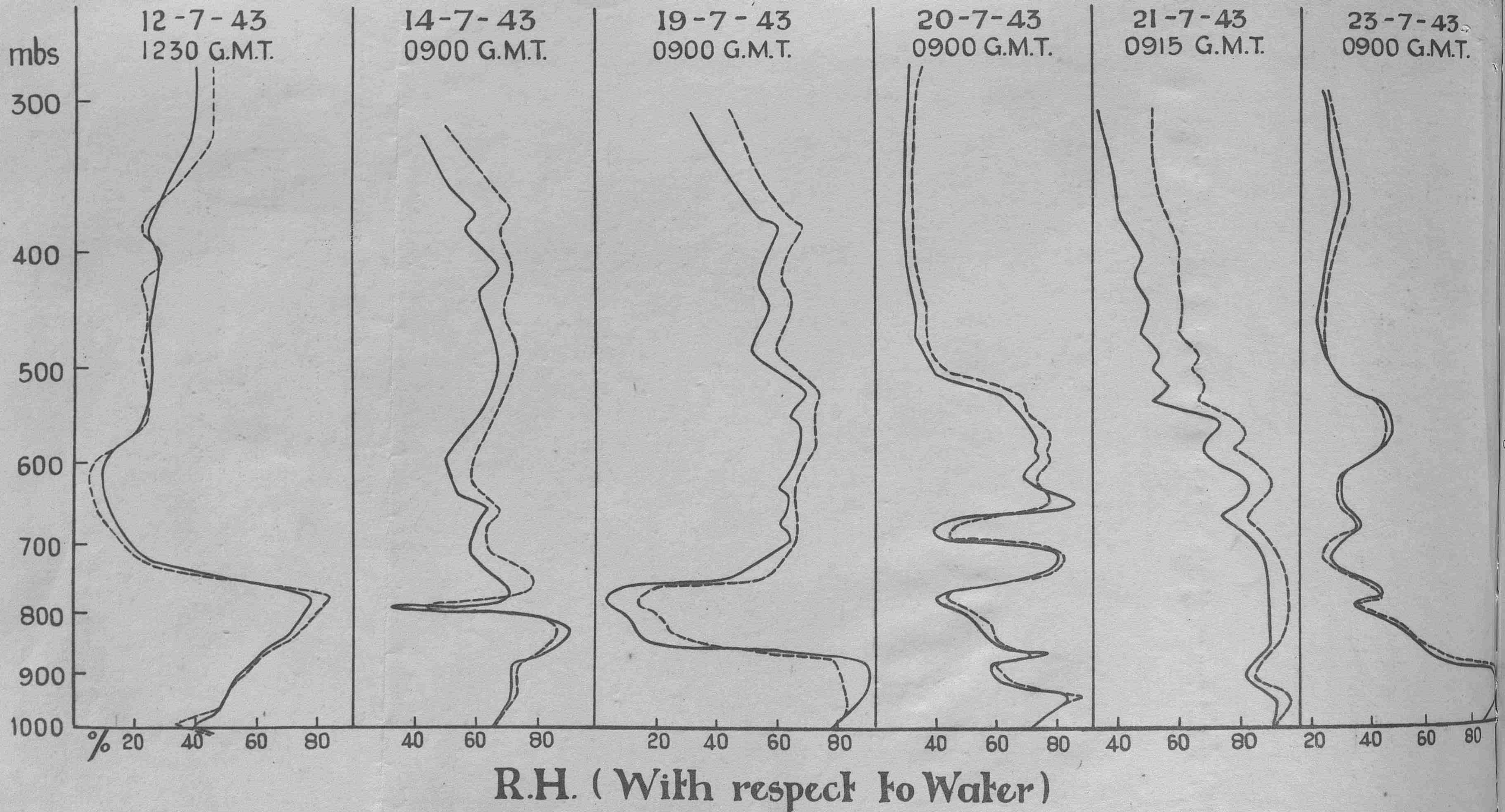
Casual errors are therefore on the average about 5% R.H. range i.e. ± 2½% ; the maximum difference found was 10% .

There are of course other sources of error which the present investigation would not reveal, the chief of which is the lag in response in rapidly changing conditions at the lower temperatures.

A temperature coefficient of the element, due to the mumetal, is not corrected for, Its neglect introduces both systematic and casual errors, the latter arising because the temperature sensitivity varies by a factor of at least three to one in different elements. The magnitude of the errors is not great however. Taking a temperature of -40°C and humidity 40% the average systematic correction to be applied is +3% , while the casual errors due to variation between elements may be as great, but are already included in the figures shown in Tab.II. The neglect of this temperature effect is thus justified.

An examination of the ascent charts shows the great faithfulness with which even small details are followed by the two elements. Many small changes are shown which would escape reporting, if not observation, in routine conditions. It is possible that some are too local to be of significance for synoptic use. It should be noticed that there is little similarity between traces here shown and those reported by Larkhill, within three hours of the ascent.

/Conclusion.


12-7-43
1230 G.M.T.
14-7-43
0900 G.M.T.
19-7-43
0900 G.M.T.
20-7-43
0900 G.M.T.
21-7-43
0915 G.M.T.
23-7-43
0900 G.M.T.
mbs
300
400
500
600
700
800
900
1000
%
20
40
60
80
R.H. (With respect to Water)

Conclusions.

The casual error of a humidity determination is shown to be in the neighbourhood of ± 2½% .

There is a systematic error, due to interaction, which may amount to 9% at 300 mbs. Its magnitude is very variable from instrument to instrument. Its origin has been located and it has been eliminated in all further calibrations so that there is no necessit to consider correction for it.

Neglect of the temperature coefficient of the humidity element is shown to be of no consequence.

The finer details in the recordings are shown to be real features of the humidity structure.

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~~Restricted~~

MRP 207
29th September,
1944.

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

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REPORT
ON AN INVESTIGATION OF SUBSIDENCE IN THE FREE ATMOSPHERE

by

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REPORT
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I. INTRODUCTION

Subsidence is usually defined as the large scale downward motion superimposed upon the horizontal currents of the atmosphere. This definition will be retained and the descending currents that occur in convective situations (e.g. between the cumulus clouds) will not be regarded as subsidence, although the physical characteristics of the air in the descending branches of the convective currents will be similar to those observed in the currents defined as subsidence.

Subsidence is known to occur in the air to the lee of mountain ranges. This kind of subsidence may be referred to as orographic subsidence. The present report deals only with subsidence in the free atmosphere, such as is independent of orographical conditions.

The aim of this investigation has been:-

(a) To investigate subsidence theoretically with a view to finding means of identifying subsiding air masses.

(b) To derive statistical characteristics of subsiding air masses.

(c) To correlate the occurrence of subsidence with the synoptic situations at sea level and aloft, with a view to finding practical rules as an aid in the forecasting of subsidence.

Some theoretical aspects of subsidence are discussed in Section II, and the results of the statistical investigation are described in Section III.

With regard to practical rules for the forecasting of subsidence only few results have been obtained. Noteworthy is the statistical result that about 80 per cent of the investigated cases of subsidence have been found to be associated with anticyclonic vorticity. Further theoretical and empirical research will be required to explain this relationship.

The characteristics of subsidence described in the following sections should be regarded as the results of a preliminary investigation which should be supplemented by detailed examination of individual cases.

In some respects the characteristics described in this report differ from those found by other investigators. The differences may, at least in part, be due to the circumstance that the material used by the various investigators represents selected cases, reflecting differences in the principles of selection. With regard to the present report it should be noted that it deals with only widespread and persistent subsidence within the British Isles. The limitations that derive from this choice are indicated in Section III, and the principal points of difference between the present and other investigations are described briefly in Section IV.

The writers wish to express their gratitude to Miss G. Beverly for the great assistance rendered in the working up of the observational material.

II. SOME THEORETICAL RELATIONSHIPS

1. Conservatism

With a high degree of approximation we may assume that the physical processes in subsiding air masses are adiabatic. Disregarding the initial phase in which saturated air may descend along the saturated adiabatic, the entire subsidence process may be regarded as dry-adiabatic, and the following quantities will be conserved* simultaneously:-

(i) potential temperature
(ii) humidity mixing ratio
(iii) potential wet-bulb temperature.

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*Potential temperature and humidity mixing ratio would not be conserved if precipitation fell through the subsiding air. The influence of evaporation from falling precipitation will not be considered here since there is no evidence that such occurred in connection with the synoptic situations investigated.

Experience shows that the vertical distribution (or the lapse rate) of these quantities assumes characteristic forms in subsiding air masses. We shall, therefore, investigate the lapse rate properties in some detail.

Let S denote a conservative property with a lapse rate given by:

$$\Gamma = -\frac{\partial S}{\partial z} \qquad (1)$$

The condition of conservatism is expressed by:

$$\frac{dS}{dt} = 0 \qquad (2)$$

and this imposes certain restrictions upon the variations of the lapse rate.

Equation (2) may also be written:

$$-\frac{\partial S}{\partial t} = \mathbb{V}.\nabla S - w\Gamma \qquad (3)$$

where $\mathbb{V}$ is the horizontal wind vector, w is the vertical velocity, and $-\nabla S$ the horizontal gradient of S.

Differentiating equation (3) partially with respect to Z, and noting that

$$\frac{\partial^2 S}{\partial z \partial t} = \frac{\partial \Gamma}{\partial t} \quad \text{and} \quad \frac{\partial}{\partial z}(\nabla S) = -\nabla\Gamma$$

we obtain:

$$\frac{\partial \Gamma}{\partial t} = \frac{\partial \mathbb{V}}{\partial z}.\nabla S - \mathbb{V}.\nabla\Gamma - w.\frac{\partial \Gamma}{\partial z} - \Gamma\frac{\partial w}{\partial z}$$

Now,

$$\frac{d\Gamma}{dt} = \frac{\partial \Gamma}{\partial t} + \mathbb{V}.\nabla\Gamma + w\frac{\partial \Gamma}{\partial z}$$

and therefore,

$$\frac{d\Gamma}{dt} = -\Gamma\frac{\partial w}{\partial z} + \frac{\partial \mathbb{V}}{\partial z}.\nabla S \qquad (4)$$

which is the relation that the lapse rate of a conservative property must satisfy.

It will be seen from equation (4) that the lapse rate Γ of a moving unit of air varies (a) on account of the vertical stretching, or shrinking, of the air column, and (b) on account of the shearing advection within the air column.

The shearing advection term vanishes (i) when there is no vertical shear of the horizontal wind, (ii) when the air is horizontally uniform, and (iii) when the shear vector is perpendicular to the gradient of the quantity in question. In any of these circumstances equation (4) reduces to

$$\frac{d\Gamma}{dt} = -\Gamma\frac{\partial w}{\partial z} \qquad (5)$$

The meaning of this equation is readily demonstrated. Consider a column of air of height $\Delta Z = Z_1 - Z_0$ and lapse rate

$$\Gamma = \frac{-\Delta S}{\Delta Z} = -\frac{S_1 - S_0}{Z_1 - Z_0}$$

Now

$$\frac{d\Gamma}{dt} = -\frac{(Z_1 - Z_0)\left(\frac{dS_1}{dt} - \frac{dS_0}{dt}\right) - (S_1 - S_0)\left(\frac{dZ_1}{dt} - \frac{dZ_0}{dt}\right)}{(Z_1 - Z_0)^2}$$

But since S is conservative

$$\frac{dS_1}{dt} = \frac{dS_0}{dt} = 0$$

and furthermore,

$$\frac{dz_1}{dt} = w_1 \quad \text{and} \quad \frac{dz_0}{dt} = w_0$$

whence

$$\boxed{\frac{d\Gamma}{dt} = -\Gamma \frac{w_1 - w_0}{z_1 - z_0}} \qquad (6)$$

If there is no wind shear, the individual column would remain vertical; therefore

$$\frac{w_1 - w_0}{z_1 - z_0} = \frac{\partial w}{\partial z}$$

and equation (6) would conform with equation (5).

As will be shown in a later paragraph, horizontal uniformity is characteristic of subsiding air masses, wherefore the shear term in equation (4) may be neglected, regardless of the wind shear. It is, therefore, appropriate to discuss lapse rates on the basis of equations (5) and (6).

2. Lapse Rate of Potential Temperature

Let θ denote potential temperature, and γ_θ the lapse rate of θ, such that

$$\gamma_\theta = -\frac{\partial \theta}{\partial z}$$

which substituted in equation (6) gives

$$\boxed{\frac{d\gamma_\theta}{dt} = \gamma_\theta \frac{w_1 - w_0}{z_1 - z_0}} \qquad (7)$$

Since θ increases along the vertical, γ_θ is negative, and the change in stability of an individual unit of air is directly proportional to the vertical stretching, or shrinking, of the layer.

3. Lapse Rate of Potential Wet-bulb Temperature

Let θ_w denote potential wet-bulb temperature, and γ_w its lapse rate. Then, as explained above,

$$\boxed{\frac{d\gamma_w}{dt} = -\gamma_w \frac{w_1 - w_0}{z_1 - z_0}} \qquad (8)$$

But γ_w may be positive or negative, depending on whether the air is convectively unstable or stable. Here, then, the sign of

$$\frac{d\gamma_w}{dt}$$

is determined not only by the stretching or shrinking, but also by the sign of the lapse rate of potential wet-bulb temperature.

4. Dew-point Depression and Amount of Descent

The relation between dew-point (T_D) and saturation vapour pressure (e) can be represented by the formula

$$\frac{1}{273} - \frac{1}{T_D} = C \log \frac{e}{e_0}$$

where e_0 is the value of e that applies at $T = 273^{\circ}A$. With sufficient accuracy* we may treat C as a constant, but are obliged to use different values of this constant for water and ice.

The humidity mixing ratio is expressed by

$$m = \frac{0 \cdot 622\, e}{p - e}$$

or

$$\frac{e}{e_0} = \frac{p\, m}{e_0 (0 \cdot 622 + m)}$$

When a unit of air moves dry-adiabatically, m is conserved, and when the pressure changes by the amount δp the dew-point will vary by the amount δT_D. From the above equations:

$$\frac{\delta T_D}{T_D^{\,2}} = C \frac{\delta p}{p}$$

Simultaneously, the dry-bulb temperature change is

$$\frac{\delta T}{T} = 0 \cdot 287 \frac{\delta p}{p}$$

whence

$$\frac{\Delta T_D}{\Delta T} = \frac{C}{0 \cdot 287} \cdot \frac{T_D^{\,2}}{T}$$

According to Shaw

$$C = 1 \cdot 629 \times 10^{-4} \quad \text{over ice, and}$$

$$C = 1 \cdot 844 \times 10^{-4} \quad \text{over water}$$

and for normal ranges of T and T_D in the free atmosphere, we obtain

$$0 \cdot 13 < \frac{\delta T_D}{\delta T} < 0 \cdot 185$$

Introducing the dew-point depression

$$\Delta T = T - T_D$$

it follows that in descending motion the dew-point depression will vary in an adiabatic process at a rate which varies from 87.0 to 81.5 per cent. of the dry-adiabatic rate, or

$$\boxed{\frac{4 \cdot 7^{\circ}F}{1000\ ft} > -\frac{d}{dz}(\Delta T) > \frac{4 \cdot 4^{\circ}F}{1000\ ft.}}$$

Now, if we assume that the subsided air originally was saturated, or nearly so, the above formula will give an approximate expression for the amount of descent that has occurred. For this purpose it is sufficiently accurate to put

$$\Delta z = 220\ \Delta T \tag{9}$$

where Δz is the amount of descent in feet, and ΔT is the dew-point depression in $^{\circ}F$.

5. Rate of Descent

It is easy to derive a formula for the vertical velocity of the air, in which only space and time variations of a conservative property enter. Since the synoptic upper air charts refer to pressure levels rather than to levels fixed in space, it is necessary to take into account the vertical movement of the pressure surfaces.

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*Shaw: Manual of Meteorology, Vol. 3, p. 238.

The following symbols will be used:

W = vertical velocity of the air

W_p = vertical velocity of the pressure surface

$\frac{\partial}{\partial t}$ = local variation at a point fixed in space

$\frac{\delta}{\delta t}$ = local variation at a point following the pressure surface

$\mathbb{V}$ = horizontal wind vector

$-\nabla$ = horizontal gradient.

In a system of co-ordinates that moves vertically with the pressure surface

$$\frac{\delta p}{\delta t} = 0$$

Therefore

$$\frac{dp}{dt} = \mathbb{V}\, \nabla p + (W - W_p)\frac{\partial p}{\partial z}$$

and also

$$\frac{dT}{dt} = \frac{\delta T}{\delta t} + \mathbb{V} \cdot \nabla T + (W - W_p)\frac{\partial T}{\partial z}$$

where $\frac{\delta T}{\delta t}$ is the temperature variation in the pressure surface.

Since the motion is adiabatic

$$\frac{1}{e} \cdot \frac{dp}{dt} = C_p \frac{dT}{dt}$$

and, therefore,

$$\frac{\delta T}{\delta t} = (W - W_p)(\gamma - \gamma_d) + \frac{A}{eC_p}\mathbb{V} \cdot \nabla p - \mathbb{V} \cdot \nabla T \tag{10}$$

With the exception of W, all terms in equation (10) can be evaluated from upper air pressure contour charts.

The first term on the right hand side is proportional to the stability of the air and the speed with which the air moves through the pressure surface.

The term containing ∇p is normally small, and vanishes when the motion is balanced. The last term represents the horizontal advection of temperature.

In many typical subsidence cases, the horizontal temperature distribution is sufficiently uniform for the advection term to be neglected; and if the motion is sufficiently balanced, the above equation reduces to

$$\boxed{\frac{\delta T}{\delta t} = (W - W_p)\gamma - \gamma_d)} \tag{11}$$

This equation will be used in a later section for evaluating the vertical velocities in some situations where the simplifications appeared to be valid.

6. Characteristics of the lower Limit of Subsidence

From the above equations, it is possible to derive a few characteristic features that will aid in the identification of subsiding air.

Experience shows that the lower limit of subsidence is well defined in all physical properties, and that there is a tendency for the base of the subsiding air to remain stationary at about 900 to 800 mb.

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the lapse rate of potential temperature decreases rapidly downward towards BB, where isothermal, or inversion conditions are found. Within the inversion, the potential wet-bulb temperature may decrease, or increase, along the vertical, according to whether the subsided air originally was convectively stable or unstable. The dew-point depression is usually very large, and reflects the total amount of descent. The dew-point depression reaches a maximum value at some distance above the base of the subsidence inversion, for the air below the inversion is usually moist, and a certain amount of mixing takes place through the inversion. In consequence of this, the amount of descent of the air near the base of the layer computed by the aid of formula (9) may be somewhat less than the actual amount.

III. STATISTICAL CHARACTERISTICS

9. Selection of Subsidence Cases

The synoptic and statistical investigation covers the period March - August 1943. This period, which was chosen at random, contains a sufficient number of subsidence cases to yield significant statistical results.

In interpretating the statistical results it is necessary to bear in mind that the cases selected represent pronounced, widespread and persistent subsidence.

The qualification "pronounced" will indicate that the subsidence had progressed to such a state that the typical features of lapse rate and vertical structure, described in the foregoing section, had had time to develop to such an extent that there could be no doubt that they had resulted from subsiding motion.

By "widespread and persistent" subsidence is understood subsidence that affected several stations at more than one time of observation.

During the period selected a number of cases of local subsidence or subsidence of short duration were found, but these cases have not been investigated. There were also a number of cases indicative of initial subsidence; but in these cases the typical features of lapse rate and vertical structure were not sufficiently definite to indicate with certainty, that they had resulted from subsiding motion. These cases, too, were eliminated.

Excluding the above-mentioned cases of local shortlived and initial subsidence, there remained 54 days within the chosen period on which widespread and persistent subsidence occurred. These cases comprised 295 ascents as shown in Table I. A list of the 295 cases is given in an Appendix.

TABLE I

Station	No. of Ascents
Larkhill	66
Liverpool	54
Penzance	43
Lerwick	28
Aldergrove	35
Bircham Newton & Downham Market	69
	Total 295

These cases have been grouped according to the circulation pattern (e.g. cyclonic, anticyclonic or straight motion, fronts, cols etc.). Through this grouping 75 individual cases of circulation patterns occurred.

It is evident that the selection of cases influences the statistical results. Thus the statistical characteristics indicated below will be more pronounced than they would have been had all cases of subsidence been included. Furthermore the frequency, or probability, of subsidence indicated below is considerably lower than it would have been had all cases of subsidence been included.

It is also of interest to note that many of the statistical results represent "one-way" relationships. Thus, for example, when it is found that about 80% of all cases of subsidence occur together with anticyclonic vorticity it does not mean that subsidence will occur in 80% of all cases when anticyclonic vorticity has been recorded.

In the statistical and synoptic investigation subsidence has been identified by the aid of the typical features of dryness, lapse rate and vertical structure described in the foregoing section. It is evident that the same features would be recorded in air that had recently completed its downward motion and been advected into the network of Upper Air Stations by the horizontal currents. There is, however, evidence to show that the statistical results refer to active, and not past subsidence, for examination of the synoptic situations showed that the air travelled with a speed considerably in excess of the speed of travel of the circulation pattern. Normally, the air would then come into the circulation pattern investigated from a system in which the presence of fronts and clouds would indicate that no subsiding motion could occur, and the typical features indicating subsidence would have developed within the circulation pattern investigated. Although it may not have been possible to eliminate cases of advection altogether, the vast majority of the selected cases would represent active subsidence, and the statistical results would, accordingly, show a strong bias for subsidence conditions.

Certain statistical results reflect conditions typical of the British Isles, and cannot be generalized. Thus, for example, the association between the occurrence of stratus and strato-cumulus and subsidence is not likely to apply to geographical regions in which the general circulation, the distribution of land and sea, etc. differ essentially from those typical of the British Isles.

10. The lower Limit of Subsidence

The frequency of occurrence of the lower limits of subsidence is shown in Table II.

TABLE II

Percentage Frequency of Lower Limit of Subsidence

Pressure Intervals	ICAN Heights (ft.)	Percentage of Frequency
At or below 950 mb.	At or below 1770	6.2
950 - 900	1770 - 3240	17.3
900 - 800	3240 - 6390	59.6
800 - 700	6390 - 9876	16.2
700 - 600	9876 - 13792	0.7
Above 600	Above 13792	0.0

It will be seen from the table that there is a distinct tendency for the lower limit of subsidence to occur in the layer from 900 to 800 mb.

It has been customary to assume that the dryness resulting from subsidence tends to find its lower limit at the top of the so-called friction layer. The frequencies shown in Table II indicate that the lower limit of subsidence is normally well above the level that is commonly understood to represent the top of the friction layer.

It will also be seen from Table II that only in about 6 per cent. of all cases the base of the subsiding layer was at or below 950 mb. In none of the situations was subsided air found lower than 40 mb. or about 1,000 feet above station level.

In some cases notable dryness has been recorded also at station level, but in these cases the dryness was due to diurnal heating, resulting in destruction of the inversion and mixing of the subsided air with the surface air.

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11. Lapse Rate at the lower Limit of Subsidence

The frequency of types of lapse rates of temperature immediately above the base of the subsiding layer is shown in Table III.

TABLE III

Frequency of Types of Lapse Rates

Type	Per Cent.
Inversion	80.3
Isothermal	11.2
Lapse	8.5

It should be noted that the frequency of occurrence of inversion conditions is higher than it would have been had all cases of subsidence been included (see para. 9).

Table IV shows the association between lapse rates of dry-bulb and potential wet-bulb temperature.

TABLE IV

Association (per cent.) of Lapse Rates of dry-bulb and Potential Wet-bulb Temperatures

Dry-bulb Temperature Wet-bulb Potential Temperature	Lapse	Isothermal	Inversion	Total
Lapse	8.5	4.7	26.8	40.0
Isothermal	1.7	2.4	13.2	17.3
Inversion	1.0	1.4	40.3	42.7
Total	11.2	8.5	80.3	100.0

The most striking feature of this table is the high proportion of cases in which there is a lapse of potential wet-bulb temperature within the subsidence inversion, indicating that the inversion air is about as often convectively unstable as it is convectively stable. Not only is the frequency of convective instability very high but also is the intensity of such instability remarkably pronounced*.

The discussion in paras. 3 and 6 showed that subsiding motion cannot change the sign of the convective stability, or more specifically, a layer that is originally convectively stable (or unstable) will remain convectively stable (or unstable), but the intensity of the stability (or instability) will increase during the subsiding motion if $\frac{\partial w}{\partial z} < 0$. There is therefore no tendency for convectively stable air to become convectively unstable, or vice versa.

In comparison with the last column of Table IV the following figures indicate the normal frequency of occurrence of the significant lapse rates within the layer from 900 - 800 mb. over Larkhill.

Potential Wet-bulb Temperatures	Normal Frequency %
Lapse	47.3
Isothermal	14.7
Inversion	38.0

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*The intensity figures are not of much interest and are therefore not reproduced here.

It will be seen that the frequency of occurrence of convective instability associated with subsidence is not essentially different from what it is normally. Furthermore, according to equation (8a), the intensity of convective instability associated with subsidence inversions must be interpreted as a result rather than a cause of, or prerequisite for, subsidence.

12. Magnitude and Depth of the Subsidence Inversion

Some additional characteristics of subsidence inversions are shown in Table V.

TABLE V

Percentage Frequencies of Depth and Intensity of Subsidence Inversions

Depth of Inversion (mb.)	Magnitude of Inversion ($^{\circ}$F.)			Total
	1-5	6-10	11-15	
10	4.3	2.6	0.4	7.3
20	13.5	6.3	0.4	20.2
30	14.3	7.2	1.3	22.8
40	10.1	5.5	1.3	16.9
50	13.9	6.7	0.4	21.0
60	2.6	1.7	-	4.3
70	2.1	0.4	0.4	2.9
80	2.1	-	-	2.1
90	0.8	-	-	0.8
100	1.7	-	-	1.7
Total	65.4	30.4	4.2	100
Mean magnitude = 4.6°F. Mean depth = 38 mb.				

It will be seen that the subsidence inversion is normally rather shallow; in 88% of all cases the depth of the inversion was 50 mb., or less. Thus, the layer characterised by negative lapse rate of temperature is very shallow, and this together with equation (7a) seems to indicate a steep gradient of the subsiding velocity through the inversion.

13. Clouds

In the majority of cases considerable amounts of cloud occurred under the subsidence inversion indicating that widespread and persistent subsidence is only rarely identifiable by clear sky*.

Table VI shows the percentage frequency of various total cloud amounts associated with widespread and persistent subsidence.

TABLE VI

Total Cloud Amount	Percentage Frequency
$0 - \frac{1}{10}$	13
$\frac{2}{10} - \frac{3}{10}$	10
$\frac{4}{10} - \frac{6}{10}$	13
$\frac{7}{10} - \frac{8}{10}$	14
$\frac{9}{10} - \frac{10}{10}$	50

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* See para. 9.

The typical subsidence clearings that are often observed occur mostly in connection with cold fronts and rapidly travelling wedges of high pressure. Such cases of shortlived subsidence are not included here.

It will be seen that in 64% of the cases the cloud amounts exceed seven-tenths.

The typical cloud forms recorded together with subsidence are shown in Table VII.

TABLE VII

Percentage Frequency of Cloud Forms

Cloud Forms	Percentage Frequency
No Low Clouds	17
Cumulus	17
Strato-cumulus	64
Fog	2
No High or Medium Cloud	59
Alto-stratus	9
Alto-cumulus; cirro-cumulus	32

It will be seen that strato-cumulus is the predominant type of cloud associated with subsidence. These clouds are invariably found below the subsiding layer, the top of the cloud sheet reaching to, and in some cases slightly above, the base of the inversion.

The characteristic feature of the subsiding air is absence of high and medium clouds, although in many cases patches of alto-stratus and alto-cumulus are observed.

14. Upper Limit of Subsidence

Unlike the lower limit, the upper limit of the subsiding air is rarely well defined. Both theory (see para. 7) and observation show that there is no sharp discontinuity in lapse rates and dew-point depression along the upper boundary of the subsiding air. Nevertheless, it was thought of interest to obtain an estimate of the frequency with which more or less marked variations indicating an upper limit of subsidence occur at various heights. The upper limit has been arbitrarily defined as "sharp" or "gradual" in the zone in which one of the following criteria were found to be satisfied:

(i) The upper limit is referred to as being "sharp" when the dew-point depression decreased upwards through a layer of 50 mb. by at least 10°F. to a value less than 10°F., provided it did not exceed 10°F. at any observed higher level.

(ii) The upper limit is referred to as being "gradual" in cases where (i) was nowhere satisfied and where the dew-point depression decreased to 5°F. or less.

The frequency distribution of the upper limit of subsidence (as defined above) is shown in Table VIII.

/Table VIII

TABLE VIII

Percentage Frequency of Upper Limits of Subsidence

Pressure Intervals	ICAN Heights (ft.)	Sharp Limit	Gradual Limit	Total
Below 800 mb.	Below 6390	0.3	0.3	0.7
700 to 800 mb.	9876 - 6390	5.4	0.7	6.1
600 to 700 mb.	13792 - 9876	11.9	0.7	12.6
500 to 600 mb.	18278 - 13792	11.9	4.1	15.9
450 to 500 mb.	20800 - 18278	2.4	2.0	4.4
400 to 450 mb.	23561 - 20800	1.4	4.7	6.1
350 to 400 mb.	26615 - 23561	1.4	2.0	3.4
300 to 350 mb.	30049 - 26615	0.3	0.7	1.0
Above 300 mb.	Above 30049	-	-	18.3
Unidentifiable				31.5

In 93 out of the 295 cases the upper limit was unidentifiable, usually owing to lack of humidity observation at higher levels. In nearly all of these cases the upper limit was very high, in the majority certainly above the 350 mb. level.

It will be seen from Table VIII that only in a minority of the cases did the subsidence terminate below the 500 mb. level, and that in such cases the limit was normally sharp*. In the remaining cases, the majority in number, the subsidence affected the major part of the troposphere above the 900-800 mb. level. In these cases the upper limit of subsidence was either gradual or unidentifiable.

15. Amount of Descent

As shown in para. 4, the approximate amount of descent may be computed from the dew-point depression on the assumption that the air was saturated, or nearly so, before the descent commenced.

Table IX shows the mean and maximum dew-point depressions recorded at various pressure levels, and the corresponding amounts of descent.

TABLE IX

Dew-point Depression (°F.) and Amounts of Descent (ft.)

Pressure	ICAN Height (ft.)	Dew-point Depression		No. of Obs.	Amount of Descent	
		Mean	Max.		Mean	Max.
950	1770	13	26	24	2860	5720
900	3240	19	51	53	4180	11220
800	6390	27	51	221	5940	11220
700	9876	27	51	271	5940	11220
600	13792	24	47	236	5280	10340
500	18278	22	47	189	4840	10340
450	20800	21	41	175	4620	9020
400	23561	20	40	151	4400	8800
350	26615	18	38	123	3960	8360
300	30049	16	35	54	3520	7700
250	33980	19	25	4	4180	5500

It will be seen that the mean as well as the maximum dew-point depressions are remarkably constant above the 900 mb. level. The maximum amount of descent is about 11,000 feet. Even at the 250 mb. level the average descent amounts to about 4,200 feet, showing that, on the average, the air that descended to this level would have originated somewhere near the 200 mb. level.

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*In many of these cases, the sharp limit was due to advection of non-subsiding air at high levels (e.g. approaching fronts)

Using the heights appropriate to the International Standard Atmosphere, and adding the mean and maximum amounts of descent to the heights of the appropriate pressure surfaces, we obtain the approximate levels at which the air commenced subsiding. These levels are shown in Table X.

TABLE X

Mean and Maximum Level (Ft.) of Origin of Subsiding Air

Level at which observed		Mean level of origin	Maximum level of origin
Pressure	ICAN Height (ft.)		
900	3240	7420	14460
800	6390	12330	17610
700	9876	15816	21096
600	13792	19072	24132
500	18278	23118	28618
450	20800	25420	29820
400	23561	27961	32361
350	26615	30575	34975
300	30049	33569	37749
250	33980	38160	39480

From this and the preceding tables it appears that widespread and persistent subsidence tends to affect the whole troposphere. The origin of the subsiding air at high levels has not been investigated in further detail.

16. Rate of Descent

Equation (11) has been used for computing the rate of descent of subsiding air masses in some cases where the horizontal temperature gradients were sufficiently small for the advection term to be neglected. For this purpose the term

$$\delta T / \delta t$$

was evaluated from the Upper Air Charts as the temperature change during 6 hours at the pressure levels concerned. The equation then gives the rate of descent per 6 hours. In individual cases the formula may give erratic values but in the majority of cases and on the average the results appear consistent.

Table XI shows, as an example, the mean rates of descent over a 36-hour period, and Table XII shows the individual values during a typical 6-hour period.

TABLE XI

Mean Vertical Velocities over a 36-hour Period

Pressure level (mb.)	800	700	600	500	400	300
(ft. per 6 hours)	-400	-590	-840	-670	-1690	-2270
(cm./sec.)	-0.6	-0.8	-1.2	-0.9	-2.4	-3.2

TABLE XII

Typical Vertical Velocities for a 6-hour Period

Pressure level (mb.)	800	700	600	500	400	300
(ft. per 6 hours)	-540	-1890	-2650	-2270	-2770	-4820
(cm./sec.)	-0.8	-2.6	-3.7	-3.2	-3.9	-6.8

The computed vertical velocities refer to the situation of 25th-26th July in which the subsidence was deep and widespread. It was found that the rate of descent decreased rapidly toward the base of the subsiding layer, was sensibly constant in mid-air and increased fairly rapidly from the 500 to the 300 mb. surface. No computation of the rate of descent was made above the 300 mb. level.

The magnitude of the rate of descent referred to above is in good agreement with that found by other authors, using different methods of determination.

Equation (7) shows that the rate of change in lapse rate of potential temperature depends primarily on the magnitude of $\partial w / \partial z$. It is of interest, therefore, to compare the observed lapse rates with the computed rates of descent. Table XIII shows the mean lapse rates corresponding to the mean vertical velocities given in Table XI.

TABLE XIII

Mean Vertical Velocity and Lapse Rates

Pressure (mb.)	800	700	600	500	400	300
W (cm./sec.)	-0.6	-0.8	-1.2	-0.9	-2.4	-3.2
Lapse rate °F./1000 ft.	2.1	2.9	3.3	3.2	4.1	4.5

It will be seen that the lapse rate is relatively constant within the zone where the rate of descent is constant, and increases in the upper layers in proportion to the increase in the rate of descent. These variations are in agreement with the theoretical deductions, and it may, therefore, be assumed that the computed rates of descent are reasonably reliable.

17. Sea Level Pressure

Let f denote the frequency of occurrence of subsidence within an interval of sea level pressure, and let F denote the frequency of occurrence of pressure within the same interval. The percentage probability of occurrence of subsidence is then equal to

$$100\ {}^{f}/_{F}$$

The selected cases of subsidence have been grouped according to intervals of 10 mbs. and the corresponding probabilities are given in Table XIV.

TABLE XIV

Percentage Frequency and Probability of Subsidence

Pressure Intervals mb.	Percentage Frequency f	Probability $100\,{}^{f}/_{F}$
< 1000	0.7	1.3
1000 - 1010	7.5	3.6
1010 - 1020	45.5	8.9
1020 - 1030	36.8	9.7
1030 - 1040	8.5	12.9
> 1040	1.0	25.0

It will be seen that the probability of subsidence increases markedly with increasing sea level pressure. Thus when the pressure is below a 1000 mb. the probability is about 1: 100, and when the sea level pressure is about 1040 mb. the probability is about 1:4.

It should be noted, however, that Table XIV refers to widespread and persistent subsidence; if all cases of subsidence had been included, all probability figures would have been considerably higher.

The charts examined are the surface weather chart, the 750, 500 and 300 mb. pressure contour charts. On the surface weather chart it was difficult to obtain reliable assessments of the shear, and the subsidence cases were grouped only according to curvature of the isobars. The results of the examination of the sea level chart are shown in Table XVII.

TABLE XVII

Percentage Frequency of Curvature of Sea Level Isobars and Subsidence

Curvature	Frequency %
Col	5
Straight	18
Cyclonic	17
Anticcyclonic	60

The upper air charts were examined according to wind shear as well as curvature of the contour lines. The association, expressed as a percentage, is shown in Table XVIII. In this table the row "anticyclonic" contains the cases in which the vorticity was definitely anticyclonic, and the row "cyclonic" contains the cases in which the vorticity was definitely cyclonic. The row "intermediate" contains the cases where curvature and shear were of opposite signs or where the flow pattern was too indefinite for assessment of vorticity.

TABLE XVIII

Percentage Frequency of Vorticity in Subsidence Cases.

Vorticity	Pressure Surfaces		
	750 mb.	500 mb.	300 mb.
Both anti-cyclonic	79	82	79
Intermediate	18	14	15
Both Cyclonic	3	4	6

The majority of intermediate cases were of anticyclonic shear with straight or cyclonically curved, contours. There is thus a very high association between subsidence and anticyclonic vorticity.

It is necessary to bear in mind that Tables XVII and XVIII represent "one-way" associations; if all cases of anticyclonic and cyclonic vorticity were examined, largely different association figures might be found for the occurrence of subsidence. Furthermore the tables refer to widespread and persistent subsidence and the figures need not be representative of the passing subsidence that occurs along cold fronts. In fact most of the cases indicated as "intermediate" represent frontal subsidence where the sign of the vorticity could not be assessed with certainty.

A cursory examination of the cases of frontal subsidence showed that most of these occurred in connection with cold fronts with a strong thermal wind in the warm air causing the warm air above the cold front surface to move faster than the front itself. In these circumstances the kinematical boundary condition would require downward motion of the warm air above the front. This point, however, requires further examination in detail.

IV. COMPARISON WITH EARLIER INVESTIGATIONS

Certain general but essentially qualitative features of subsidence are already well known and are discussed in standard text books* (1, 2). We shall therefore not refer to them here. There is, however, much additional literature /

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* Reference to literature will be found at the end of this section.

touching on the subject (3, 4, 5), and in particular on the use of potential wet-bulb temperature as a means of identifying subsiding air masses, and also on the relation between subsidence and horizontal divergence in the free atmosphere. Some investigators have endeavoured to compute the horizontal divergence by the aid of pilot balloon observations, and from these computations the vertical velocities have been inferred. The discrepancies between the results obtained by the various investigators may be due partly to the selection of cases for the investigation and partly to the methods used. It is therefore of interest to compare the methods as well as the results derived.

23. Method of Conservatism

Hewson (6) first developed the method of determining the air trajectories by identifying the air masses by the use of potential wet-bulb temperature. Doubt has been thrown on the validity of Hewson's results by Bleeker (7). Bleeker's argument is that, strictly speaking, potential wet-bulb temperature is conservative with regard to evaporation, and potential pseudo-wet-bulb temperature is conservative with regard to all adiabatic processes. As regards the precise definitions of these quantities the reader is referred to (2). Bleeker's criticism is, however, considered to be invalid, for the minor difference between the two kinds of potential wet-bulb temperature could not influence the results derived by Hewson. What has been said above of Hewson's investigations applies also to that of Longley (8).

Hewson's method consists in considering potential wet-bulb temperature as a conservative property and identifying the travelling air masses at two, or more, consecutive instances. In the present investigation this method has not been used, for on account of the small horizontal extent of the network of upper air stations, identification of individual units of air of sufficiently small dimensions was only rarely possible. However, the methods of identification used in the present investigation are entirely based on the principle of conservatism, as set out in Section II, using all conservative elements. In principle, therefore, there is no difference between the method used by Hewson and Longley and that used by the present writers.

24. Computation of Divergence

Theoretically it is possible to use the equation of continuity and observations of the wind distribution for determining:

(i) the local pressure variations, and
(ii) the vertical velocity.

Whether or not accurate results can be obtained depends entirely on the accuracy of the wind observations.

In order to obtain an estimate of the observational accuracy required for computing the pressure variations, we consider a vertical column of air from sea level to the top of the atmosphere with a base of 100 km. x 100 km. and one of the vertical sides facing the wind. To obtain a decrease in pressure from 1000-999 mb. during one hour it is necessary to increase the distance between the windward and the lee sides of the column by 1:1000, or 100 m., to which corresponds a wind variation from the one side to the other of 3 cm./sec. Therefore, to obtain an estimate of barometric tendencies of the order of 1 mb./hour, the winds at all levels and stations must be measured with an accuracy of some few mm./sec.

To obtain an estimate of the accuracy required for the computation of vertical velocity we consider a slice of air 1 Km. deep and horizontal cross-section 100 Km. x 100 Km. The order of magnitude of the vertical velocity in subsiding air is 2-4 cm./sec. to which corresponds a horizontal wind variation from one side of the air column to the other of about 2-4 m./sec. To obtain such wind differences with sufficient accuracy it is necessary that the wind observations at all levels and stations be accurate to about 20-40 cm./sec.

/Even

APPENDIX

SELECTED ASCENTS USED IN THE STATISTICAL INVESTIGATION

Station abbreviations:

ALD - Aldergrove
BCN - Bircham Newton
LAR - Larkhill
DMK - Downham Market
LER - Lerwick
LIV - Liverpool
PEN - Penzance

Date	Time	Station	Top of Ascent if below 200 mb.
Mar. 15	0000	PEN	
"	0000	LAR	
"	0624	BCN	400
"	0700	LAR	
"	1200	LAR	
"	1200	LIV	
"	1206	BCN	400
Mar. 18	0824	BCN	400
"	0000	LAR	
"	0000	LIV	
"	0000	PEN	
"	0706	ALD	600
"	0600	LAR	
"	0600	LER	
"	0606	LIV	
"	1154	ALD	400
"	1200	BCN	400
"	1200	LAR	250
"	1200	LIV	
"	1700	BCN	400
"	1800	LER	210
Apr. 4	0000	LER	
"	0000	LIV	
"	0000	PEN	
"	0600	LIV	
"	0600	BCN	400
"	0700	LER	
"	1200	LAR	
"	1200	LIV	
"	1200	PEN	
Apr. 5	0000	DMK	
"	0000	LIV	
"	0600	LIV	
"	1200	LIV	200
"	1736	BCN	400
Apr. 6	0000	PEN	
"	0600	LAR	
"	0600	LIV	
"	1200	LIV	
"	1200	PEN	
Apr. 8	0000	DMK	
"	0000	LIV	
"	0100	LAR	
"	0100	LER	
"	0600	DMK	310
"	0600	LAR	280
"	0600	LER	
"	0612	BCN	400
"	1800	LER	

/Apr. 9

Date	Time	Station	Top of Ascent if below 200 mb.
Apr. 9	0000	PEN	
"	0600	LAR	
"	1200	LAR	
"	1200	LIV	
"	1200	PEN	
Apr. 10	0000	LAR	
"	0000	LIV	
"	0000	PEN	
"	0600	BCN	400
"	0600	LIV	
"	1200	LAR	
"	1200	LIV	
"	1212	BCN	400
"	1200	PEN	
"	1748	BCN	400
Apr. 11	0000	PEN	
"	0100	LAR	
"	0548	ALD	400
"	0600	LAR	
"	1148	ALD	400
"	1200	LAR	
"	1200	PEN	
Apr. 17	0000	LAR	
"	0000	LER	
"	0600	LAR	
"	0600	LIV	
"	1200	ALD	400
"	1800	LER	
Apr. 20	0100	LIV	
"	0600	LAR	
"	0600	LIV	
"	1200	LIV	
"	1200	PEN	
Apr. 21	0000	LAR	
"	0000	PEN	
"	0600	LAR	
"	1200	BCN	400
"	1200	LIV	
"	1800	BCN	400
Apr. 22	0000	LER	
"	0800	LER	470
"	1200	PEN	
Apr. 25	0000	LAR	
"	0000	LER	
"	0000	PEN	
"	0600	LAR	
"	1200	LAR	
"	1200	PEN	
May 12	0000	DMK	
"	0000	LAR	
"	0700	LAR	
"	1200	BCN	400
"	1800	BCN	400
May 15	0000	DMK	
"	0000	LAR	
"	0000	LER	
"	0200	LIV	
"	0600	LAR	
"	1200	LIV	
May 16	0000	LAR	
"	0000	PEN	
"	1200	LIV	
"	1200	PEN	
"	1800	LER	

/May 22

Date	Time	Station	Top of Ascent if below 200 mb.
Aug. 3	0000	DMK	350
"	0000	LAR	
"	0000	LIV	
"	0600	LAR	
"	0600	LIV	
"	1200	BCN	400
"	1200	LAR	
"	1200	LIV	
"	1200	PEN	
"	1748	ALD	400
"	1748	BCN	400
Aug. 4	0000	LIV	200
"	0000	PEN	
"	0100	DMK	200
"	0600	BCN	400
"	0600	LAR	
"	1154	BCN	400
"	1200	LAR	
"	1800	BCN	400
Aug. 5	0000	LAR	200
"	0600	LAR	
"	0600	LER	
"	1200	LAR	200
"	1800	LER	200
Aug. 8	0000	DMK	
"	0000	LAR	280
"	0000	PEN	
"	0600	LAR	
"	1200	ALD	400
"	1200	PEN	
"	1742	ALD	400
Aug. 9	0554	ALD	400
"	0600	BCN	400
"	0600	LAR	
"	1148	ALD	400
"	1200	LAR	
"	1800	ALD	400
"	1800	BCN	400
"	1800	LAR	
Aug. 10	0000	DMK	
"	0000	LIV	
"	0000	PEN	
"	0554	ALD	400
"	0600	LAR	
"	0600	LIV	
"	1200	LIV	
"	1200	PEN	
"	1300	LAR	
"	1800	BCN	400
"	1800	LAR	
"	2100	BCN	400
Aug. 11	0000	DMK	
"	0000	LAR	
"	0000	LIV	
"	1200	LAR	
"	1200	PEN	
"	1800	LAR	

AIR MINISTRY

MRP 209
20th October 1944.

METEOROLOGICAL RESEARCH COMMITTEE

Portable Cloud Searchlight (Ceiling Projector).

by

O.M.Ashford.

1. Introduction. The standard Meteorological Office cloud searchlight, described in the Meteorological Observer's Handbook, 1942, pp. 75-77, was intended primarily for permanent installation at fixed stations. For use by mobile units it suffers from the following serious disadvantages:- (a) Weight. The weight, not including cable and transformer, is 150 lbs., while the packing case weighs another 60 lbs. (b) Mounting. On account of the weight of the equipment and the necessity for maintaining an accurate level, the searchlight has to be mounted on a solid concrete base.

(c) Power Supply. The lamps are usually 250 watt, 24 volt. The electrical installation calls for the services of a skilled electrician. The difficulties mentioned under (b) and (c) frequently lead to delays of a week or more between the arrival of the searchlight and the completion of the installation.

To meet the needs of meteorological units attached to mobile R.A.F. units, there was obviously a requirement for a light portable searchlight which could be installed and maintained very easily, and which could detect clouds up to at least 3000 feet.

2. Description. The instrument that has been developed to meet these requirements is illustrated in figures 1 and 2. It consists essentially of a standard motor headlamp mounted on a simple galvanised steel base, a length of cable on a light drum, a relay, an alidade and three 6 volt car batteries.

(a) Headlamp. This is a Lucas type CFSF 150/ lamp fitted with a clear front glass. The reflector is a $7\frac{1}{2}$" diameter pressed brass parabolic mirror, silver plated and polished. The bulb has a V filament and clear glass, and is nominally rated at 12 volt 36 watt. The total weight of this assembly including relay and base is only 19 lbs. For transport the lamp is packed in a special wooden box weighing some 5 lbs.

(b) Cable and drum. 1000 feet of Ducel cable is supplied on a drum which is fitted with a robust switch. The total weight of this part is 35 lbs.

(c) Relay. A Londex relay with two sets of 5 m.m. silver "make" contacts in parallel is mounted inside the housing A on the searchlight base. (This provides remote control of the light). The switching coil consumes about 1 watt at 18 volts, and the main circuit is capable of carrying 10 amps.at 18 volts.

(d) Alidade. The portable alidade or clinometer consists of an engraved alimunium plate A, mounted freely on a sight bar B and weighted by a brass weight C so that it remains in a steady position as the sight bar is pointed at the spot on the cloud. The height of the cloud for a baseline of 1000 feet between the alidade and the searchlight is indicated

/against

against a pointer.

(e) Batteries. Any 18 volt supply capable of giving 4 amps will suffice. Three standard 6 volt car batteries weighing 35 lbs. each are supplied when required, but in many cases it is possible to utilise batteries already available at the station which have been supplied for other purposes.

3. Installation. The searchlight base is provided with 3 short spikes B which should be pushed well into the ground. The instrument can then be levelled by adjusting two brass levelling screws (not shown); the level is indicated by a small spirit level C on a length of chain, which should be placed on the lamp glass for this purpose. If this operation is carried out with ordinary care, the beam will be within about half a degree of the vertical. After levelling, the upper base plate is clamped to the lower plate by a wing nut.

After the battery and remote control plugs are inserted in the relay housing and the batteries connected up, the cable drum is rolled away from the searchlight until all the cable is unwound. This provides a baseline of 1000 feet between the searchlight and the drum. The searchlight can be switched on and off by means of the switch D on the drum and the observer should stand beside the drum when taking readings with the portable alidade.

The whole installation can be completed easily in half an hour.

4. Operation and Maintenance. After switching on, the alidade is pointed at the reflected spot on the cloud using the luminised ball sights. When it is pointing in the correct direction, it is clamped in position by pressing the thumb on the spring clamp D, and the height of the cloud is read off direct on the engraved scale.

Apart from ensuring that the cover glass is kept clean the searchlight itself requires very little maintenance. The chief thing is to look after the batteries properly, i.e. to keep the terminals clean and charge them regularly.

The bulbs are held in a collar and its position can be adjusted to obtain the best focus. This point requires careful attention when a new bulb is fitted, but the whole procedure can be done in a few minutes.

5. Performance.

(a) Beamwidth. Owing to the V shape of the bulb filament, the beam is approximately oval in section. The minimum beamwidth (the angular width of the section of the beam with an intensity at the edges one tenth that at the centre) is approximately 3° and the maximum is nearly 4°.

For comparison the beamwidth of the standard cloud searchlight is about 3°.

(b) Intensity. The bulbs are nominally 12 volts 36 watts, but by operating them at 18 volts the power is increased to 70 watts and the axial intensity of the beam is increased from some 116,000 candles to over 400,000 candles. The average life of the bulbs at 18 volts is roughly 8 hours, which allows for several months normal usage.

The corresponding figures for the standard searchlight with a 250 watt bulb are:- axial intensity of the beam of the order of 2,000,000 candles,

average life of bulb 25 hours.

(c) Tests on Clouds. When tested against a standard cloud searchlight, the beam is noticeably fainter, but the spots obtained are very nearly as good.

For clouds up to 5,000 feet there is very little to choose between the two searchlights, but above that the extra power of the standard searchlight is an advantage.

(d) Operational Tests. Preliminary reports from meteorological units in the field, indicate that the equipment is proving to be satisfactory from an operational point of view.

6. Conclusion.

The most serious disadvantages of the instrument in its present form are the necessity for removing rain water which has collected on the glass and the ageing of the silvered mirror. For instruments produced under more leisurely circumstances both these troubles could be readily overcome, the first by fitting a sloping glass cover to the lamp as in the standard searchlight and the second by rhodium plating the mirror.

The instrument described above was designed in the Instruments Branch of the Meteorological Office with valuable assistance from Messrs. J.Lucas, Ltd., and Messrs. Best and Lloyd, Ltd.

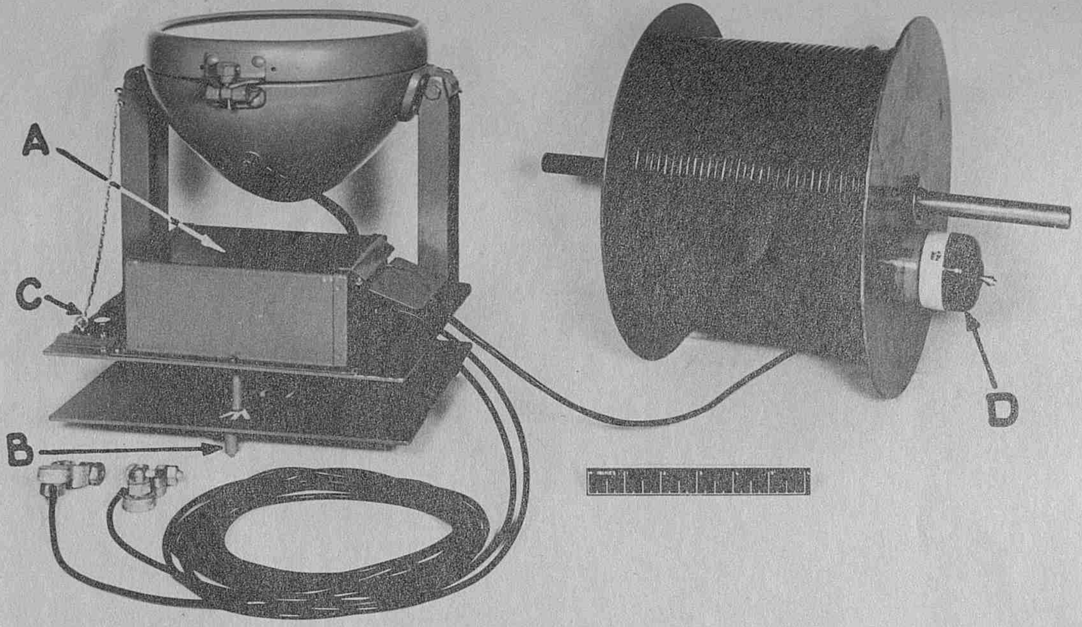
A
C
B
D

Fig. 1.

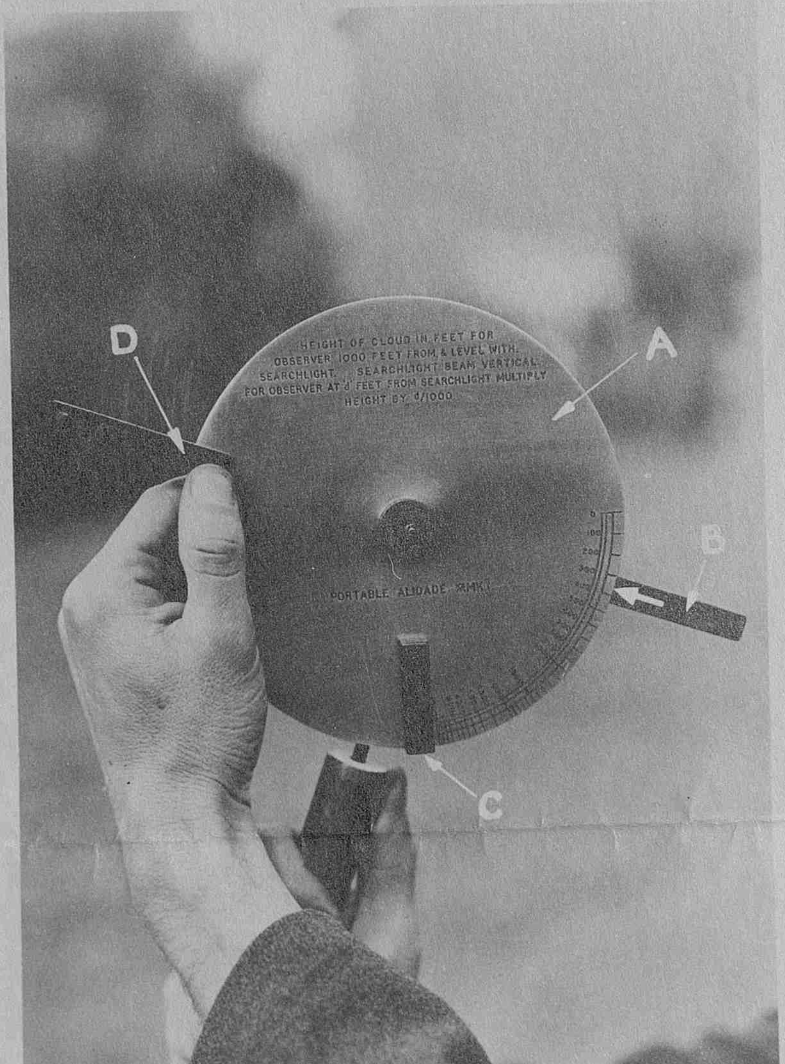
D
A
HEIGHT OF CLOUD IN FEET FOR OBSERVER 1000 FEET FROM & LEVEL WITH SEARCHLIGHT. SEARCHLIGHT BEAM VERTICAL. FOR OBSERVER AT d FEET FROM SEARCHLIGHT MULTIPLY HEIGHT BY d/1000
0
100
200
300
PORTABLE ALIDADE MK I
B
C

Fig. 2.

AAF Weather
Weather Central Library

AIR MINISTRY

M.R.P. 213.

METEOROLOGICAL RESEARCH COMMITTEE

Amendment.

Classification cancelled (or changed to) Unclassified
by authority of APR 305 12 Oct. 1946
by John Petrik, Maj. date 16 Apr. 1959

Section 1. Paragraph 3.

Delete from "(a) to conform...." to end of paragraph, and substitute:- ✓

"in various parts of the picture to conform with the information available about (i) humidity either obtained directly or deduced from reports of contrails, (ii) cloud and (iii) temperature and lapse rate changes with time. It will be seen that they are also broadly the vertical movements needed to maintain the equation of continuity in the section under consideration, having regard to the change in the disposition of troposphere and stratosphere from fig. 1 to fig. 2."

This time a deep low passed W. to E., its centre going almost along the Arctic Circle and then again a bridge of high pressure built up across the British Isles.

Diagrams showing the temperature distribution in a vertical section roughly perpendicular to the general circulation were prepared by plotting the datafor Penzance, Larkhill, Liverp ool, Aldergrove, Wick and Lerwick. Observations at 20h. each day were available except for Aldergrove at which 17h. was the nearest hour of observation and Wick at which 7h. next* day was the nearest available. Fig. 1 shows the structure for March 9d. 20h. and Fig. 2 that for March 10d. 20h.

The horizontal arrows are intended to show the transfer of air to north or south or more strictly towards the left or the right across the surface run of the isobars. They have been computed from the radio wind data. The vertical arrows show the upward or downward movement of air required (a) ~~to conform with the horizontal convergence or divergence and (b) to conform also to such information as is available about (i) humidity either obtained directly or deduced from reports of contrails, (ii) cloud and (iii) temperature and lapse rate changes with time.~~

The vertical section of Fig. 1 points to two circulations, one below and one above the 400 mb. level (about 23,000 feet which as it happens is near the level of approximately constant air density). In the lower one the air movement is upward at the north end of the picture, southward in the levels from 10,000 or 15,000 feet, to 23,000 ft. downward at the south end of the picture and northward along the ground. The upper circulation gears into this lower circulation so that it runs in the other direction, i.e., downward at the north end, southward in the upper troposphere and lower stratosphere and upward at the south end. Presumably at very high levels there may be a return northward circulation.

Fig. 2 shows the further development of these circulations a day later. In the lower cirulation the northern part is now occupied by colder (Polar) air, but the movement is still upwards as evidenced by CuNb tops at Wick at about the 20,000 feet level and a relative humidity of 98% at Lerwick at the same level. On the other hand the relative humidity at Larkhill in the 15,000 to 20,000 feet level has come down from 95% at 6h. to 55% at 13h. and to 10% at 20h. on the 10th March, indicating descent of air.

2. Effects of the Upper Circulation.

The upper circulation (in the upper troposphere and stratosphere) is better developed in Fig. 2. Without (so far as can be seen) any real change

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* Used in this paper in the sense of initial and final stages as recorded at a fixed station; as a rule a pressure system is developing as well as travelling and with a narrow area of upper air stations it is not usually easy to distinguish the travel from the development effects in a moving system.

G61.2

~~RESTRICTED~~

AIR MINISTRY

MRP 213
18th December 1944

METEOROLOGICAL RESEARCH COMMITTEE

The Vertical Circulations in relation to the Momentum and Density Distribution in Cyclones.

by A. H. R. Goldie.

Summary.

The vertical section between cyclone and anticyclone gives evidence of two circulations, one below the 400 mb. level - roughly the level of quasi-constant air density - and the other above this level, running in opposite senses as if geared into one another. The consequential effects on stratosphere temperatures and tropopause features are illustrated; the characteristic distributions of specific momentum in the initial and final stages* of depressions and in anticyclones are shown to be in accord with the distributions of air density in these respective conditions.

1. Vertical Structure - 9-12 March, 1943. The two circulations.

The first period selected for examination is 9th-12th March 1943, a period with a strong westerly circulation across the British Isles; within this time a deep low passed W. to E., its centre going almost along the Arctic Circle and then again a bridge of high pressure built up across the British Isles.

Diagrams showing the temperature distribution in a vertical section roughly perpendicular to the general circulation were prepared by plotting the data for Penzance, Larkhill, Liverpool, Aldergrove, Wick and Lerwick. Observations at 20h. each day were available except for Aldergrove at which 17h. was the nearest hour of observation and Wick at which 7h. next day was the nearest available. Fig. 1 shows the structure for March 9d. 20h. and Fig. 2 that for March 10d. 20h.

The horizontal arrows are intended to show the transfer of air to north or south or more strictly towards the left or the right across the surface run of the isobars. They have been computed from the radio wind data. The vertical arrows show the upward or downward movement of air required (a) to conform with the horizontal convergence or divergence and (b) to conform also to such information as is available about (i) humidity either obtained directly or deduced from reports of contrails, (ii) cloud and (iii) temperature and lapse rate changes with time.

The vertical section of Fig. 1 points to two circulations, one below and one above the 400 mb. level (about 23,000 feet which as it happens is near the level of approximately constant air density). In the lower one the air movement is upward at the north end of the picture, southward in the levels from 10,000 or 15,000 feet to 23,000 ft., downward at the south end of the picture and northward along the ground. The upper circulation gears into this lower circulation so that it runs in the other direction, i.e., downward at the north end, southward in the upper troposphere and lower stratosphere and upward at the south end. Presumably at very high levels there may be a return northward circulation.

Fig. 2 shows the further development of these circulations a day later. In the lower circulation the northern part is now occupied by colder (Polar) air, but the movement is still upwards as evidenced by CuNb tops at Wick at about the 20,000 feet level and a relative humidity of 98% at Lerwick at the same level. On the other hand the relative humidity at Larkhill in the 15,000 to 20,000 feet level has come down from 95% at 6h. to 55% at 13h. and to 10% at 20h. on the 10th March, indicating descent of air.

2. Effects of the Upper Circulation.

The upper circulation (in the upper troposphere and stratosphere) is better developed in Fig. 2. Without (so far as can be seen) any real change of air supply the tropopause at Larkhill has greatly sharpened up and the temperature has gone down from -76 to -95°F. Moreover at Bircham Newton at 7h. on 11th persistent contrails began above the tropopause and continued to 7,000 feet into the stratosphere, then becoming intermittent. It is difficult, having regard to these two facts, to avoid the conclusion that the stratosphere air here (as well as the upper troposphere air) has in fact recently ascended and is saturated.

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* Used in this paper in the sense of initial and final stages as recorded at a fixed station; as a rule a pressure system is developing as well as travelling and with a narrow area of upper air stations it is not usually easy to distinguish the travel from the development effects in a moving system.

The temperature of -95°F. at the tropopause and the steep lapse rate from 28,000 feet to 40,000 feet are consistent with the whole of the air in this region having been supplied from about the 25,000 feet level, i.e. the air in this part of the high having been pushed into the high mainly originally at around the 25,000 feet level. The lapse rate and the shape of the tropopause correspond also exactly to those noted in A.R.C. paper No. 5092 as being characteristic conditions for the occurrence of persistent contrails.

Fig. 3 is produced to show in detail the changes in the region of the tropopause at Larkhill in the anticyclonic area before, during and after the passage of the northern depression. The actual tropopause level does not change much, but the temperature is brought down considerably, the lapse rate is steepened to about the saturated adiabatic from about 25,000 feet up to the tropopause and the lower stratosphere temperatures are lowered considerably - all in keeping with upward air motion in the upper troposphere and the loss of air from high levels in the stratosphere,

Fig. 4 shows the changes taking place at the same time in the upper air structure at Lerwick, but there are complications here which can be more easily explained by reference to the time graphs of Fig. 5. The last mentioned shows that at Lerwick the air temperature at 10,000 feet and also at 20,000 ft. fell gradually until about 10d.18h. and then after a sharp drop (a discontinuity) resumed its gradual fall at the 20,000 feet level. These gradual falls are the result of the upward air movement, the sudden drop marks the drawing in of a new and colder air supply from more northerly latitudes.

The four graphs of Fig. 4 show the upper air structure at Lerwick for 9d.20h., 10d.13h., 10d.20h. and 12d.5h., respectively. The first is the original relatively warm air with sharp tropopause. The second, 10d.13h., i.e. before the influx of new air, shows the effect of upward motion at the lowest levels from surface up to some 15,000 feet and the effect of downward motion in all levels above about 23,000 feet. The tropopause has been flattened away and the lower stratosphere temperature raised and (incidentally)"no contrails" are reported. (So far as the writer can see it is typical of these subsided stratospheres that there are no contrails unless possible short ones and no high cloud). By 10d.20h. the invasion of colder air up to nearly the 20,000 feet level has taken place and the tropopause is of Type II. By 11d.20h. the temperature at the 400 mb. level goes down by 15°F. as compared with that at 10d.20h.; this is consistent with the continued rise of air in this region from about the 16,000 feet level to the 22,000 feet (400 mb.) level.

At Lerwick the lower stratosphere is still warmer at 10d.20h. than at 10d.12h. and the humidity at 25,000 feet is now only 10% as against the former 55% but it cannot be said with certainty that at this and higher levels it is still the same kind of air as formerly. Probably however it is actually the same air as at 12h. (and as on the previous day) because the wind direction from 20,000 feet upwards remains identical and the speed almost identical, whilst below 20,000 feet the speed decreased notably with the change of air and the direction veered by 10° to 20°. Indeed there is every indication that the tropopause allocated officially to the ascent of 10d.20h. (18,200 feet) is actually also now the top of the indrawn polar air; since however the air originally above about 23,000 feet has been subsiding, it is possible that by 20h. all the original upper troposphere air had already flowed away horizontally, leaving only stratosphere air above. Finally the curve for 12d.5h. shows the upper air structure at Lerwick after the depression has passed and before another has arrived. It may be that this curve represents the undisturbed condition for the polar air, but it is significant that all through the sequence of these events there seems to have been little change of temperature at (and also probably above) 45,000 feet.

Fig. 5 shows that at the 200 mb. level the temperature even at Lerwick changes continuously. There is no indication of a new air supply. It is to be noted particularly that at all the levels discussed the circulation above the cyclone remains cyclonic and that above about 45,000 feet the circulation is virtually nil.

3. Transfer of the Pressure Field upwards.

It will be useful now to reconsider the results given in MRP 163 in relation to the present picture. Briefly the former results showed that initially the air towards the centre of the active cyclone is definitely of lower density than that in the surroundings of the cyclone; also that characteristically the cyclone develops first as a whirl of maximum momentum in the lower levels of the troposphere, that the tropopause is lowered and that later the momentum is transferred to the upper half of the troposphere with raising of the tropopause. It was presumed in M.R.P. 163 that this transference of momentum would be effected by the rising air, but in view of the more detailed examination of the vertical components of the circulations in the present paper this point can be further examined.

Though it is possible that the transfer of momentum from lower to higher levels is accomplished partly by the rising air, it now seems to be definitely indicated as taking place mainly by change in the pressure field in the higher levels. The 400 mb. isobar slopes decidedly more steply from high to low in Fig. 2 than in Fig. 1. This is because the air below it in the high has become warmer as a result of subsidence; the air below it in the low has become colder both by reason of ascent and by reason of colder air being drawn into the centre of the low. On the other hand the air above the 400 mb. level has in the high become colder by ascent and in the low warmer by descent. In brief then, the density conditions for a cyclonic circulation have been practically levelled away in the lower half of the troposphere, but have been transferred in accentuated form to the upper troposphere and stratosphere, where previously they scarcely existed. This fits in exactly with the changes in specific momentum described in M.R.P. 163 as between the initial and final stages of the depression.

It is to be noted that this result would not be achieved by occlusion as commonly described i.e. by the lifting of the air in the warm sector by cold air running under it. On the contrary it seems to be an essential part of the actual mechanism that there should have been a reversed vertical circulation in the upper atmosphere by means of which the depression's stratosphere is brought down and warmed adiabatically, whilst the anticyclone's stratosphere and upper troposphere are raised and cooled. It is to be noted that this explanation, with the difference that the demarcation between the two circulations was placed by him at the mean tropopause level rather than, as it now seems it should be, at about the 400 mb. level, was put forward by W.H. Dines twenty years ago as a possible mechanism to account for the mean statistical relations between temperature and pressure in the upper air.

4. Dynamical and Advectional Changes on December 4-5, 1942.

As it happens, the present writer was responsible in 1923 for putting forward advection of polar and tropical stratospheres as being a possible explanation of the stratosphere temperatures found above cyclones and anticyclones. It is very difficult in general to separate the changes in air temperature in the high atmosphere which are being produced as a result of dynamical effects in the depression from the changes which are simply the result of advection of new air. The depression of December 4-5, 1942, the specific momentum of which was considered in M.R.P. 163, section 10, can be used to obtain evidence up to a point, since the centre passed across the British Isles in about the latitude of Aberdeen. Information is thus available from Lerwick to N. of the depression, from Wick and Aldergrove not far from the centre and from Liverpool, Larkhill, Penzance and other stations to S. of the depression. The data available are plotted in cross section in Figs. 6 and 7.

In connection with the 8h. data for Aldergrove used in Fig. 7, it is necessary to point out that in the Upper Air Supplement no tropopause is allocated to this ascent; it is described as "not reached". It is however not credible, having regard to data of other stations and to the sequence of data around this time, that the tropopause at Aldergrove on the morning of December 5th was higher than the 37,000 feet reached on this occasion. Everything indeed points to it as having been at about 22,500 feet where there was a very marked inversion. Consideration of the data on a T.ϕ or Hertz diagram shows that to change from the temperature distribution of Fig.6 in the Aldergrove region to that of Fig. 7 involves bringing the air formerly at 185 mb. down to the 200 mb. level , i.e. down by 1800 feet, (a pressure rise of 15 mb.), the air formerly at 220 mb. down to the 250 mb. level, i.e. down by 3,000 feet and the air at the 250 and 300 mb. levels down to the 300 and 350 mb. levels, i.e. down by about 4,000 feet. The total change at tropopause level was about 6,000 feet. Thus the peculiar temperature distribution here is again considered to be in accord with the dynamical lowering of the tropopause to the relatively low level shown in Fig. 7.

Unfortunately there are some gaps in the temperature data but the isotherms shown in Figure 6 and 7 have been reached after a careful consideration of all the available data, including the wind data and it is thought that they are approximately correct. The successive positions of the tropopause are shown in Fig. 8.

Coming to the possibility of advection, there is, in this example, no reason to suppose that in the region of Wick or Aldergrove, the air was of more "polar" origin than at the surrounding stations of Lerwick, Larkhill etc., but certainly the tropopause in the Wick-Aldergrove region became much lower than it was originally in that region or than it was at any time at Lerwick or Larkhill during the passage of the depression; and afterwards it rose again to rather above its initial level. It is thus difficult to avoid the conclusion that the principal part of the changes near the centre was in this case dynamical, not advectional. Incidentally this depression is shown on the D.W.R. as occluded during the whole period of time here discussed.

5. General Deductions.

If the dynamical explanation here put forward is valid for cyclones it would almost inevitably point further to the difference between polar and tropical stratosphere temperatures and tropopause heights as being themselves a long term dynamical balance effected across the globe from the equator to the pole by the average planetary circulation, as suggested in M.R.P. 163.

The difference between the cyclonic and anticyclonic stratospheres and their tropopause structures and tropopause heights would thus be explained as the dynamical effect of their local circulation, complicated usually by advection of the larger scale latitude differences which are continually being built up by the average planetary circulatinn. Type II tropopauses would usually be those which were being or had recently been pulled down, and the stratospheres above them would be very dry. Type I tropopauses, where very exaggerated, would be recently formed anticyclonic and would have conditions of saturation or almost saturation below them and perhaps even also above them.

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On the matter of air density it is worth recalling that W.H. Dines, reasoning from British statistics (which on the average relate to dying cyclones) could not admit density differences as the primary cause of the vertical circulations (referred to in the last paragraph of §3) which he visualized as being needed to account for the statistical results. But the variation of momentum with height in the initial stages of depressions (as shown in M.R.P. 163) points precisely to the density differences required to initiate the circulations; and the developed stratosphere conditions illustrated in the present paper give the density distribution needed for the final momentum distribution and for Dines' statistical results. The characteristic distribution of momentum with height in anticyclones, (i.e. with a maximum in middle levels, indicating a warm lower troposphere and a cold upper troposphere) is also seen to be the necessary counterpart in the high to the vertical circulations as now described.

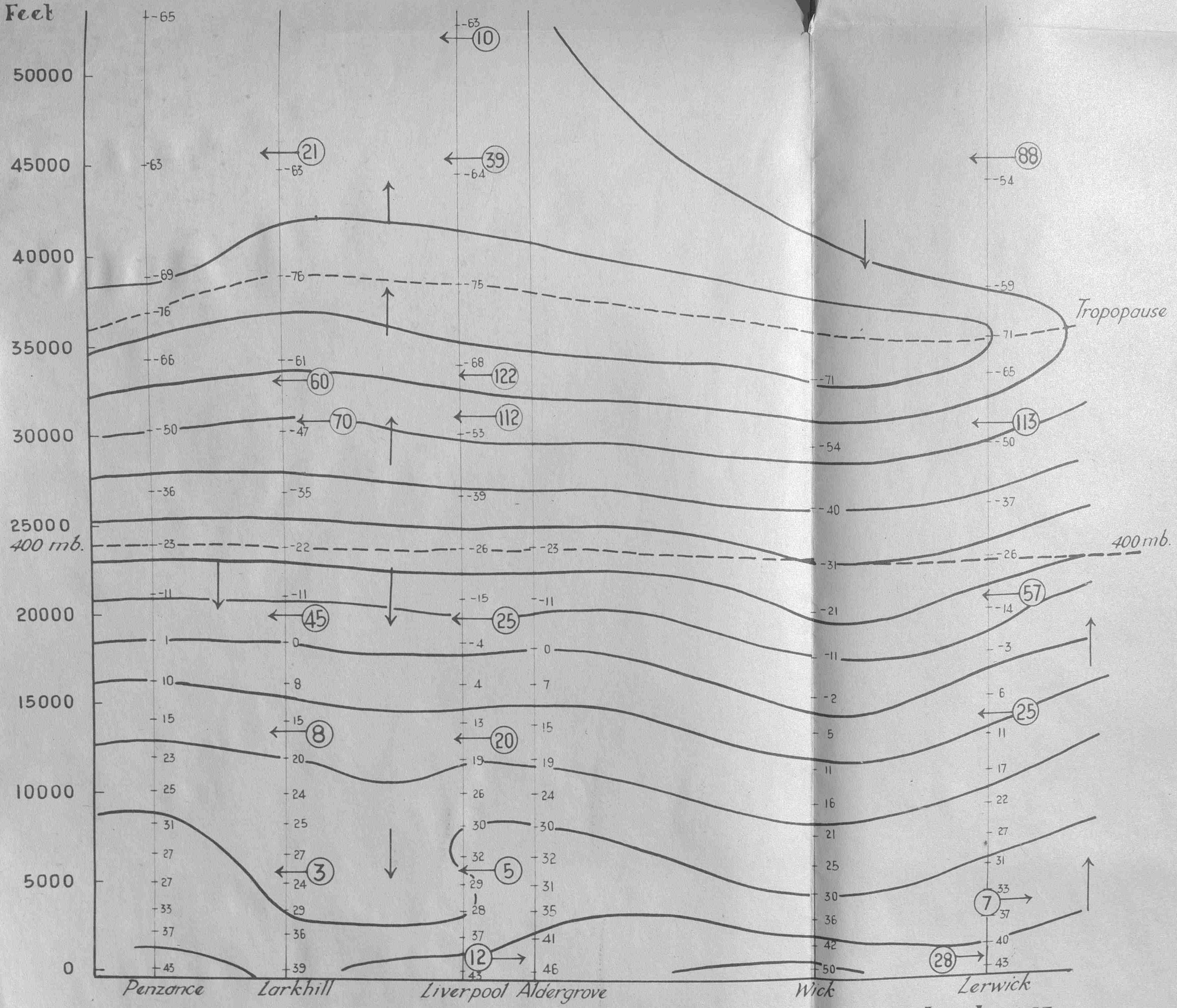

FIG. 1 - TEMPERATURE DISTRIBUTION, MARCH 9d. 20h., 1943

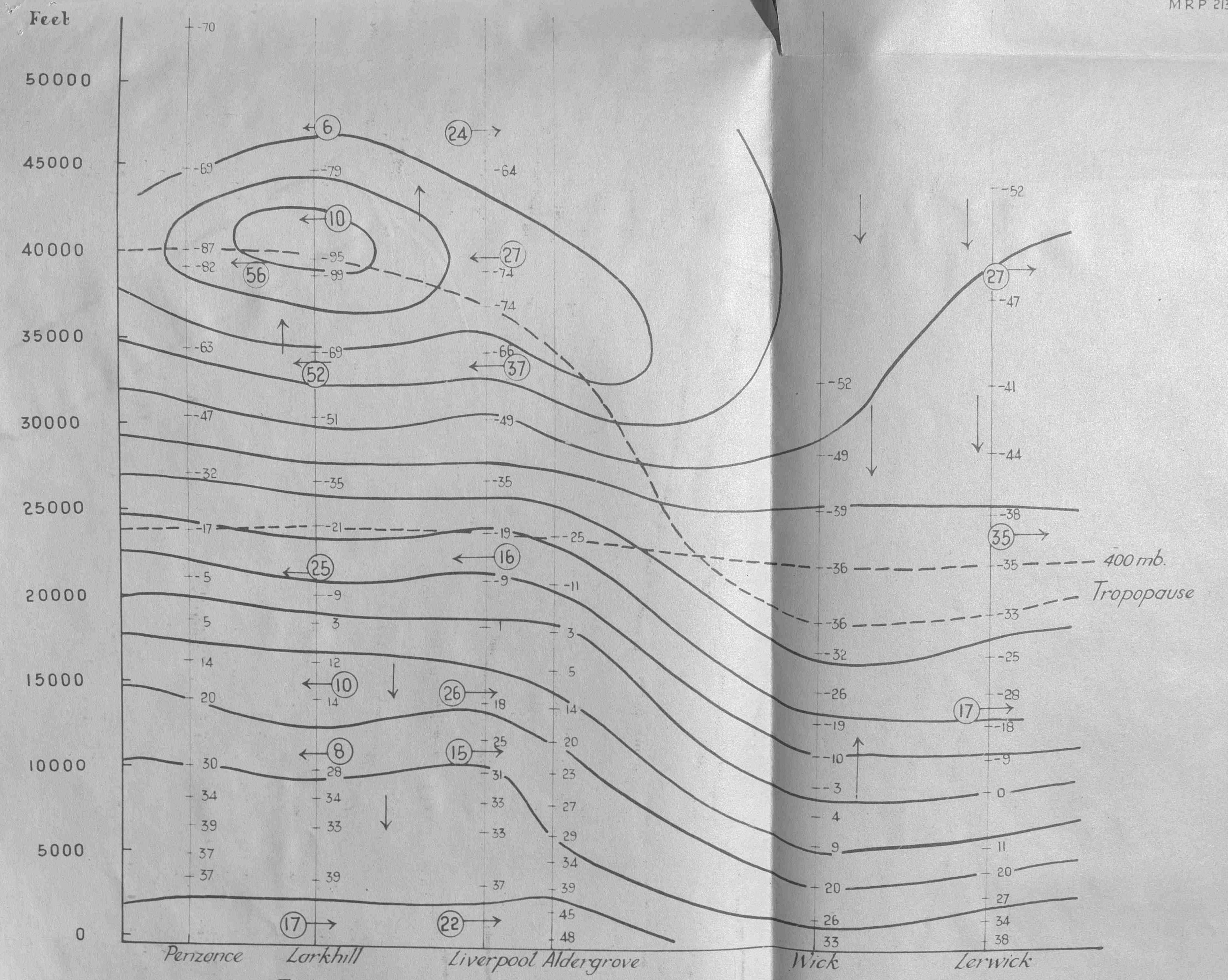

FIG. 2 - TEMPERATURE DISTRIBUTION, MARCH 10 d, 20 h., 1943

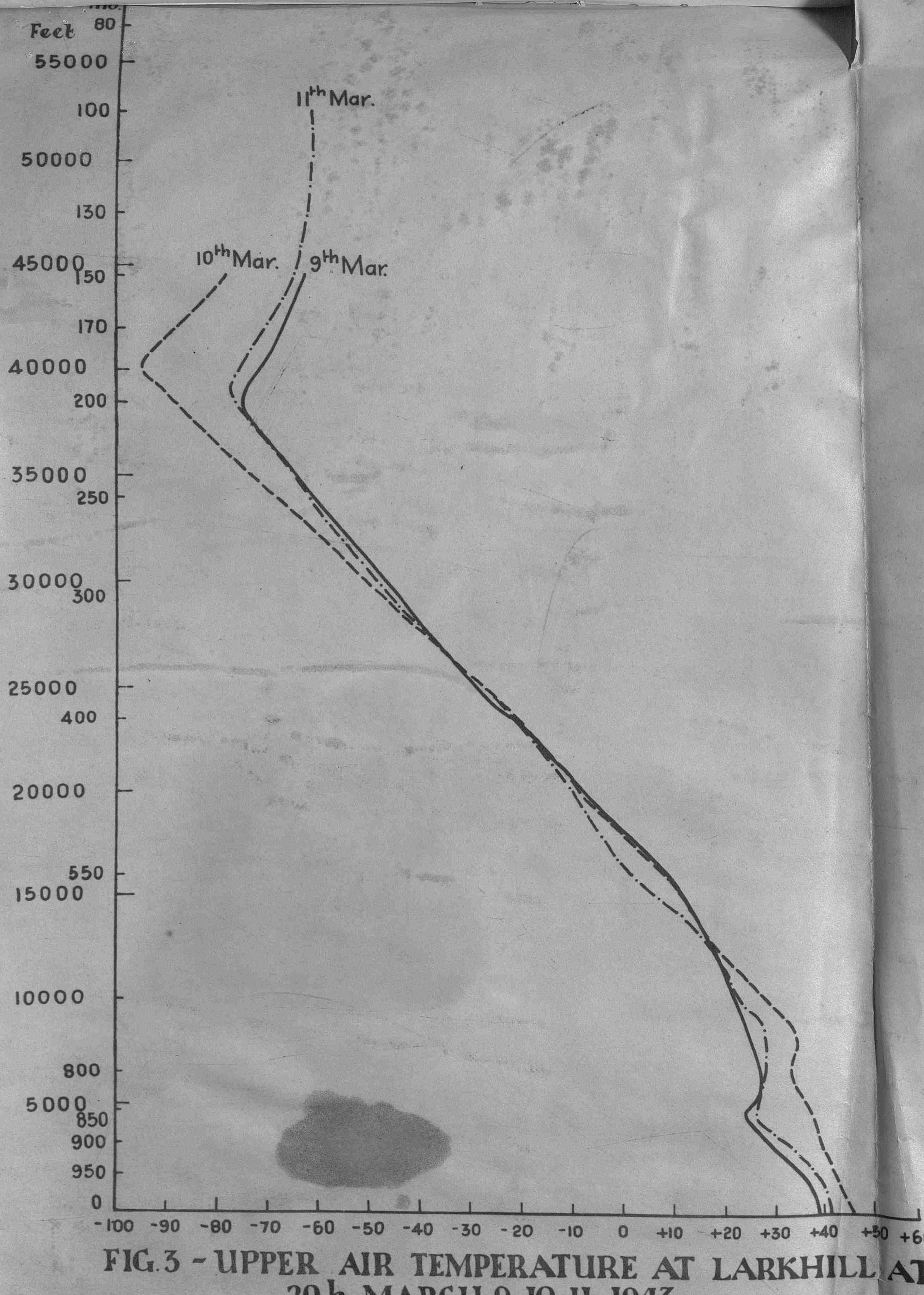

FIG. 3 - UPPER AIR TEMPERATURE AT LARKHILL AT 20 h. MARCH 9, 10, 11, 1943

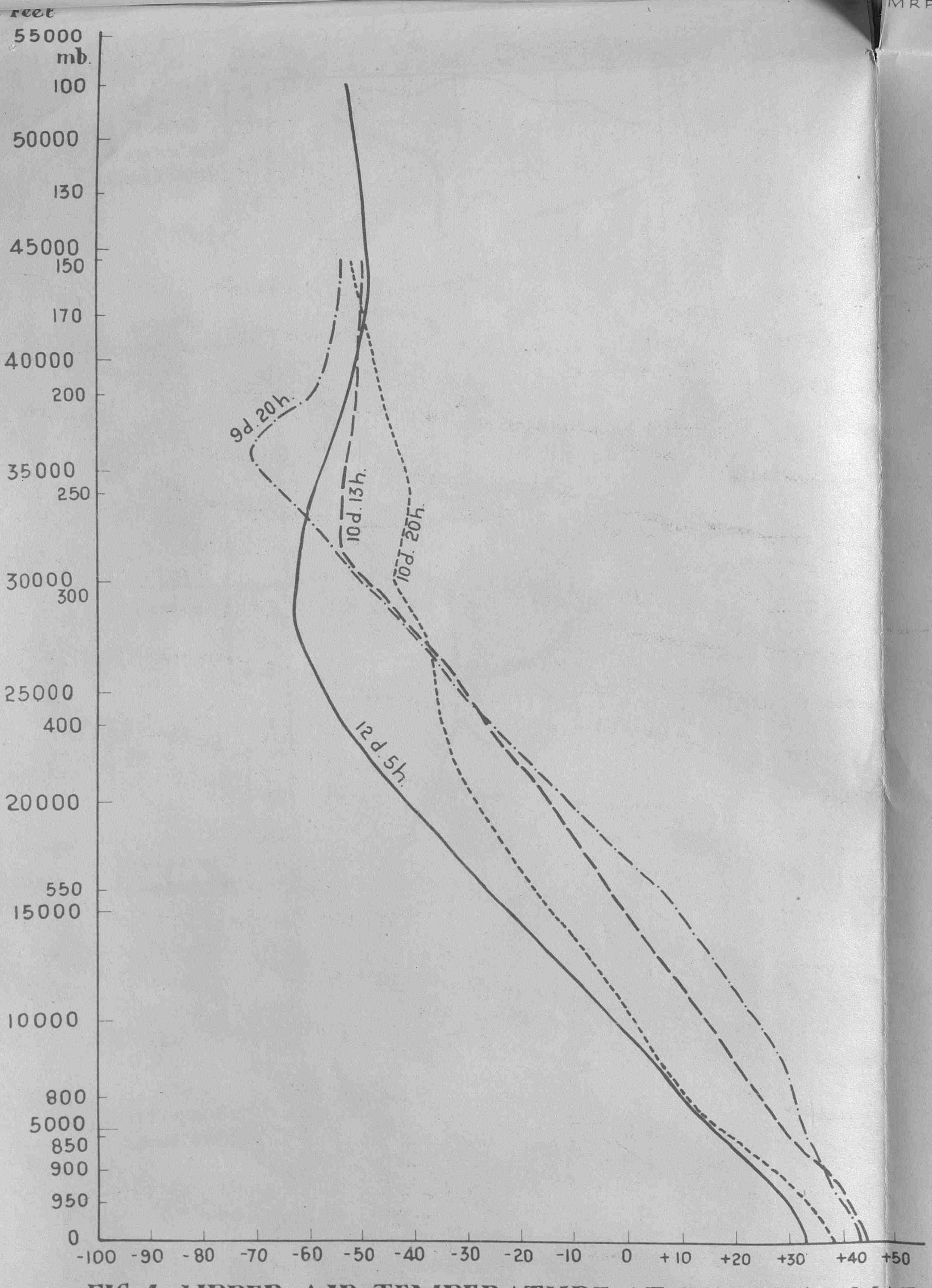

FIG. 4 - UPPER AIR TEMPERATURE AT LERWICK MARCH 9 - 12, 1943

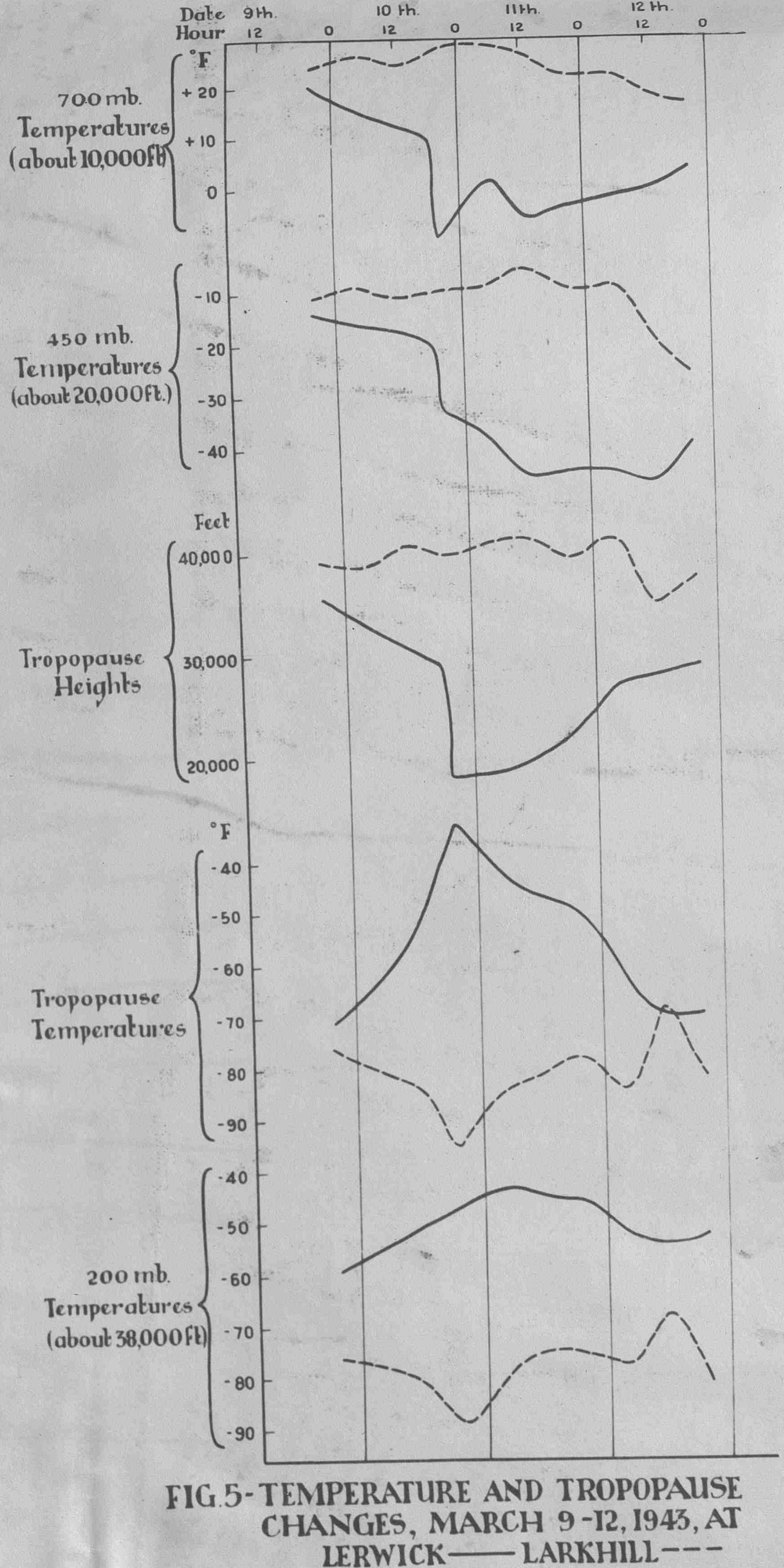

FIG. 5 - TEMPERATURE AND TROPOPAUSE CHANGES, MARCH 9-12, 1943, AT LERWICK —— LARKHILL - - -

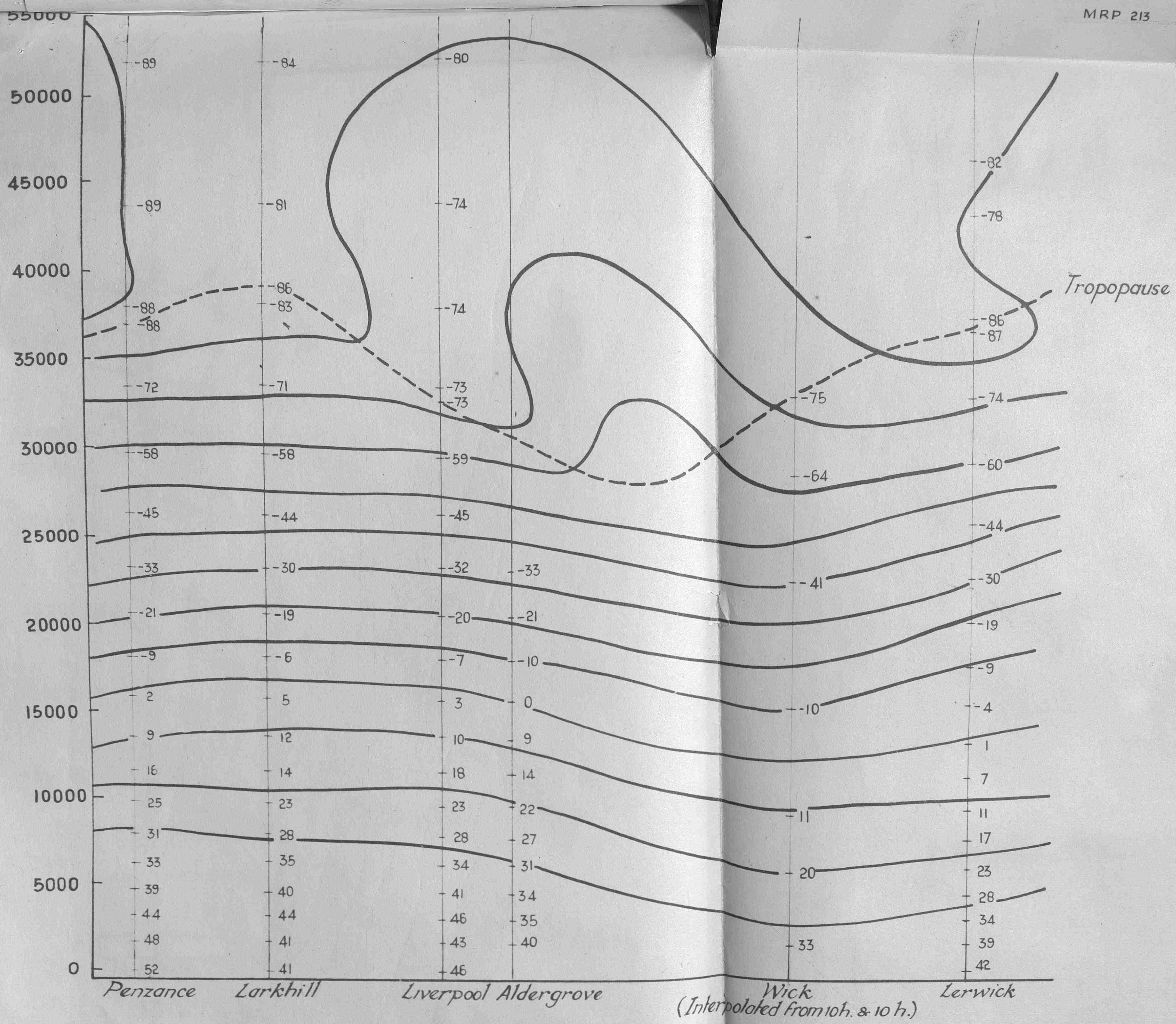

FIG. 6 - TEMPERATURE DISTRIBUTION, DEC. 4 d. 20 h., 1942

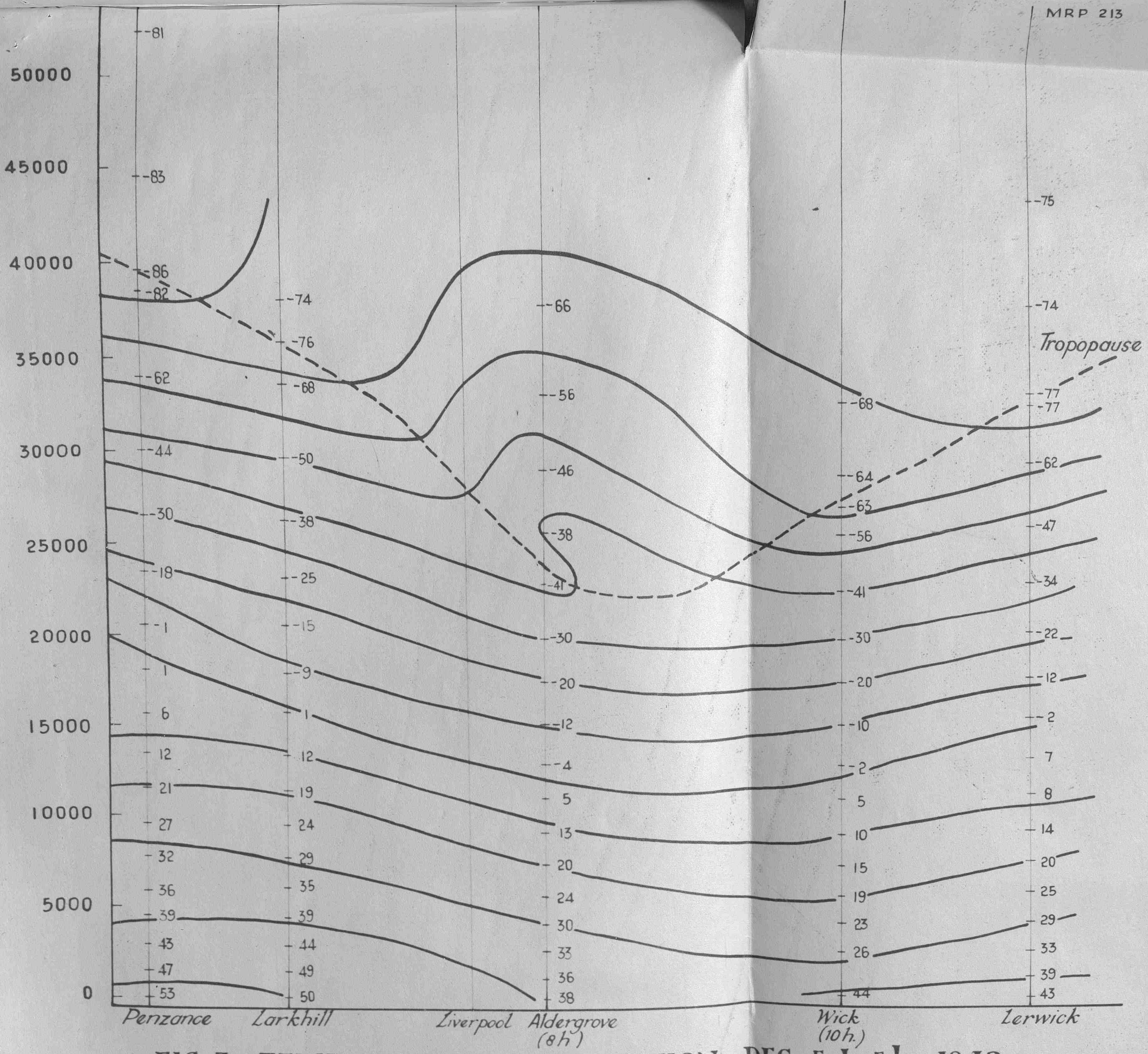

FIG. 7- TEMPERATURE DISTRIBUTION DEC. 5d. 5h., 1942

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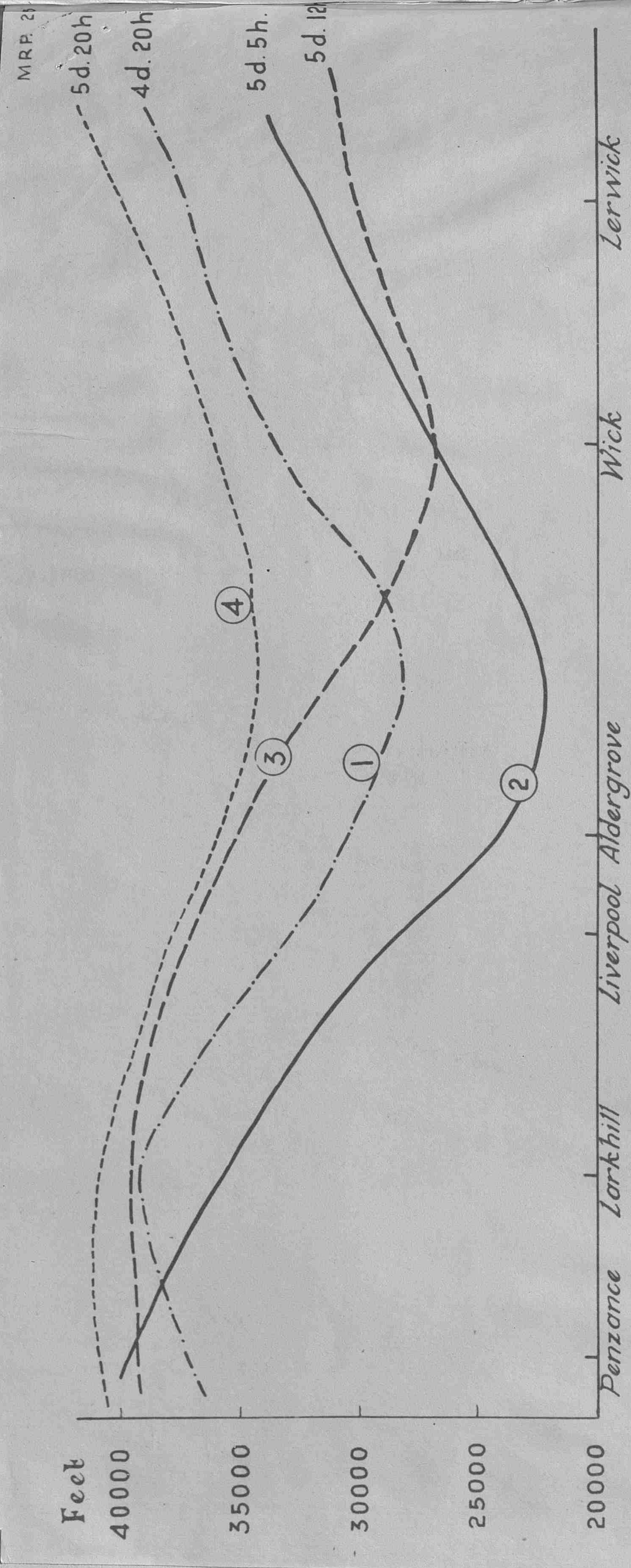

FIG. 8 - CROSS SECTIONS OF TROPOPAUSE DEC. 4-5, 1942

AIR MINISTRY

MRP 218
7 February 1945.

METEOROLOGICAL RESEARCH COMMITTEE

THE CONDENSATION AND SUBLIMATION OF WATER VAPOUR AT LOW TEMPERATURES.

Summary by G.M.B. Dobson of Report

by

B. Cwilong.

(Note:- This work was done by Mr. B. Cwilong for the Meteorological Office. His full report runs to 67 pages with many photographs and is too long to circulate to the Meteorological Research Committee. Below I give a condensed account of the report for the information of the Committee.

G.M.B. Dobson.)

Classification cancelled (or changed to) Unclassified
by authority of MRP 305 12 Oct '46
by JP date 16 Apr 59

1. Origin of the work.

When work was begun on dew-point hygrometers for use at very low temperatures it was found that frequently the deposit was in the form of dew although the temperature was far below the freezing point. Moreover it was often found extremely difficult to distinguish with certainty between a deposit of dew and one of hoar frost. Experiments were therefore made with different surfaces to try to find one on which the deposit was always hoar frost if the temperature was below the freezing point. It was also found that in most cases the surface of the thimble of the hygrometer could be cooled one or two degrees C. below the true hoar frost-point before any deposit was formed.

2. First Experimental arrangements.

In the first instance the surfaces under test were formed on the flat end of a thick copper bar, the lower end of which was cooled by placing it in a Dewar vessel of liquid air or solid CO_2 and acetone. A heating coil was wound round the upper part of the bar and the temperature was measured by a thermo-junction inserted in a small hole ending just under the test surface. The upper end of the copper bar carrying the test surface was usually enclosed by a cylindrical Dewar vessel through which dry air was passed. A microscope was arranged so that the surface could be examined by eye with a magnification of between x 50 and x 500 or the surface could be photographed. The object of using the cylindrical Dewar vessel to enclose the test surface was to keep down the inflow of heat when it was desired to surround the test surface with cold as well as dry air. The test surfaces were formed either by soldering a small piece of the metal onto the top of the bar or by evaporating various metals onto it in a vacuum. In most cases two different surfaces were formed on two halves of the end of the bar for comparison.

3. First experimental results.

Since zinc and SiO_2, like ice, crystallise in the hexagonal form, it was first thought that one of these might be suitable for the surface of the thimble, as ice might be expected to form on them very easily. In the case of a freshly made zinc surface this was in fact the case and hoar frost, rather than dew, nearly always formed if the dew-point was below 0°C. Unfortunately after a few days, even if kept in a closed vessel the zinc soon lost this property, though ice was always formed on it more easily than on most other surfaces.

Since evaporated SiO_2 is supposed to condense as a crystal rather than as a glass, SiO_2 surfaces were formed in this way. However it was found extremely difficult to form ice on such a surface when it is new, and dew, rather than hoar frost, forms even at quite low temperatures. Tests were also made in which a metal surface was lightly covered with very fine SiO_2 dust which was expected to "seed" the surface. This again had no effect. Indeed the SiO_2 surface formed by evaporation was one of the least favourable for the formation of ice, being surpassed in this respect only by a freshly cleaved surface of mica on which it was found impossible to form ice at any temperature down to at least -125°C., the deposit always appearing as dew, though freezing later.

Other metal surfaces tested have properties intermediate between those of freshly deposited zinc and freshly split mica. Polished surfaces always show more tendency for the deposit to form as dew than do surfaces formed by evaporation in a vacuum. On aluminium in particular the deposit always tends to form as dew.

As all these surfaces become old they apparently become contaminated and their properties in respect to the sublimation of water vapour become more nearly alike, being intermediate between the two extremes. The result is that no surface is specially suitable for the thimble of a hoar frost point hygrometer since either dew or hoar frost may form on any of them and there is always some supercooling before the first trace of deposit begins to form. For this reason it is absolutely necessary in using a hygrometer of this type at low temperatures to produce a deposit on the thimble by sufficient cooling and then to find the temperatures at which this deposit is in equilibrium with the water vapour in the surrounding air.

Several other interesting facts were noticed in the course of these experiments:-

For any surface which has become "old" as mentioned above, if the dew-point is between 0°C. and about -20°C the surface has to be cooled below the dew-point and only then does dew (without any sign of hoar frost,) form in microscopic droplets. If cooling is continued the drops grow and coalesce so that a film of liquid water is formed and finally the film freezes over in a flash. If the dew-point is between about -20°C and -40°C deposition starts much as before but as the drops grow they freeze on coalescence though remaining as spherical particles of ice. If the dew-point is below about -40°C. a few specks of ice are generally formed somewhat before the thimble has been cooled to the dew-point (though, of course, well below the frost-point). If cooling is continued the number of ice specks increases slightly but they are never numerous. If the temperature of the thimble falls below the dew-point dew is deposited between the ice specks. If the frost-point of the air is very low, - about -80°C., - the deposit consists of a uniform slight bloom, often difficult to see unless the light and viewing conditions are good.

If a deposit of metal formed by evaporation in a vacuum is so thin that light is easily transmitted through it, then water vapour does not easily sublime onto it. This may partly account for the phenomenon very frequently observed that if the two halves of the test surface are coated with different metals there is usually no deposit of either water or ice along a line about a millimetre wide which separates the two halves.

4. Second experimental arrangements.

The experiments described above had shown the very great effect of ageing of all surfaces, presumably due to contamination from the surrounding air. It was thought that a knowledge of absolutely clean, uncontaminated surfaces might lead to valuable results. An apparatus was therefore set up in which both the evaporation of the metal surfaces and later deposition of pure water on them, could be performed wholly in a vacuum without ever exposing the surface to air. Two metals were generally evaporated onto the test surface, a screen being placed over half the surface during the evaporation of the second metal. Water vapour was supplied from ice in a side tube enclosed in a cooling bath. (When evaporating the metals the side tube was placed in liquid air).

5. Results with the second apparatus.

Again it was found that on a surface of zinc a deposit of hoar frost started to form when the surface was not more than a few tenths of a degree below the frost-point. Cadmium behaves very much as zinc. With other metals considerable supercooling is necessary before any deposit forms and again the supercooling is greatest on aluminium. As a typical example the results for a gold surface, formed as before by evaporating the metal, are given:-

Evaporated Gold.

Temperature of water or ice in the side tube.	Temperature of surface for first deposit.	Nature of deposit.
+15°C.	+14.9°C	Dew
- 5°C.	- 6.0°C	Dew
-15°C.	-16.7°C	Ice
-23°C.	-25.1°C	Ice
-40°C.	-43.8°C	Ice

The general conclusion reached with this apparatus is that the results with freshly formed surfaces in the first set of experiments represent fairly closely those for clean metals, and no further information of special interest for the construction of a frost-point hygrometer for use at low temperatures, was obtained.

6. A Study of Sublimation Nuclei in Air.

It will be remembered that in the earlier part of the work it was shown to be surprisingly difficult to form ice by the sublimation of water vapour on a cold surface, provided that the temperature of the surface was higher than about -40°C. even though the surface had been freely exposed to the atmosphere. This indicated that there were very few "active" sublimation nuclei on the surface and therefore probably also few in the air. On the other hand the condensation of water vapour to form dew indicated that there were - as is well known - very many condensation nuclei both on the solid surface and in the air. By some curious chance - possibly owing to the difficulty of the subject - extremely little work has been done on sublimation nuclei in the atmosphere.

Mr. Cwilong suggested that he should investigate this matter as opportunities arose between other work, by repeating some of C.T.R.Wilson's experiments at low temperatures. It will be remembered that C.T.R.Wilson showed that while dense clouds were formed on expanding fresh air with quite small supersaturations, as the air became cleaner, greater supersaturations became necessary and in really clean air no condensation occurred unless the expansion ratio exceeded 1.25 when a rain-like condensation occurred, while if the ratio exceeded 1.38 a cloud-like condensation occurred.

The apparatus was similar in principle to that used by C.T.R. Wilson. The cloud chamber was bent over, being connected to the piston chamber at the top, so that it could be surrounded by a cooling bath. The cloud chamber was made large (350 c.c.) in order to increase the probability of getting nuclei in it.

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A major difficulty was to distinguish ice particles from water drops with certainty. This was overcome by using supercooled water in the bottom of the chamber as an indicator, since if one minute ice crystal fell into it the whole immediately froze, and as the walls of the chamber were also coated with supercooled water there was no chance of an ice crystal escaping detection.

The results obtained with this apparatus were remarkable and may be summarised as follows:-

(i) Results with well cleaned air.

(a) If the minimum temperature in the cloud chamber at the end of the expansion was above -35°C. only water droplets were formed and all the results were the same as if the cloud chamber had been well above freezing point.

(b) If the minimum temperature at the end of the expansion fell to -35.0°C. a few ice particles were formed along with a water fog and the supercooled water froze after a delay of between one second and a few minutes, - the delay presumably being due to the time taken for the first ice particle to fall into the supercooled water.

(c) If the minimum temperature at the end of the expansion fell to -35.5°C. or lower the supercooled water always froze immediately showing that very many ice crystals were formed. As the minimum temperature fell further below -35°C. the proportion of ice crystals increased and finally a fog of ice crystals was formed.

(d) The amount of supersaturation at the end of the expansion plays no part in determining whether ice particles or water drops are formed, the only factor being whether the minimum temperature falls below -35°C. or not.

(e) Provided that the minimum temperature falls below -35°C. ice particles are formed in thoroughly clean air even though the expansion ratio is much below the value of 1.25 which must be exceeded at higher temperatures if water drops are to be formed. It is not certain whether there is any definite minimum expansion ratio necessary for the formation of ice particles.

(f) If the air in the cloud chamber be ionised by exposure to X rays before the expansion, many more ice particles are formed but there is no change in the critical temperature of -35°C.

(g) If the cloud chamber is surrounded by a very cold bath so that the minimum temperature on expansion falls below about -80°C., the ice appears as a number of rather large grains. At still lower temperatures the number of grains becomes smaller and below about -120°C. there is practically no sublimation.

(ii) Results with atmospheric air.

(a) If ordinary atmospheric air be used in the cloud chamber the critical temperature of -35°C. is raised to about -27°C. and becomes much less sharp. If the air be grossly contaminated by tobacco smoke the the value rises to -23°C.

(b) If the experiment be made at very low temperatures the same effect is found as under i(g) indicating that all sublimation nuclei are very "inefficient" at these very low temperatures.

The fact that sublimation occurs in clean air with an expansion ratio much below 1.25, provided that the minimum temperature is below -35°C., shows definitely that the process by which solid ice particles are formed is quite different from that by which water drops are formed, and that, for example, water droplets are not first formed and then freeze.

Exactly similar results were obtained if the air in the cloud chamber were replaced by helium or carbon dioxide.

The exact figure of -35.6°C. for the critical temperature in clean air was obtained by using a triple cloud chamber connected to one and the same piston chamber, each chamber being cooled to a slightly different starting temperature which would, of course, give similarly different minimum temperatures. The temperatures of the baths used were then chosen so that the supercooled water in one or two cloud chambers froze while that in the others did not. By bracketing the temperatures in this way a very exact value was found.

The supercooled water used in these experiments as an indicator was prepared by distilling water at a low pressure without any bubbling, (which would cause droplets to be carried over). The distilled water was then boiled for a long time until the greater part had been evaporated. Samples of water prepared in this way have definite supercooling points which vary with the samples but remain constant in any one sample for a long time. Samples can be prepared which can be readily supercooled to -10°C. or lower. When the starting temperature of the cloud chamber had to be below about -10°C. the method of detecting ice by supercooled water could not be used. However it was found that if the lower part of the cloud chamber was strongly illuminated, any ice particles, which might be formed in a water fog grew rapidly in size as they fell through the supersaturated air and could be seen glittering in the beam of light. Naturally this method is not so certain for detecting ice crystals as the supercooled water but it proved satisfactory when the former could not be used.

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WEATHER SERVICE
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AIR MINISTRY

MRP 219
15th February 1945.

METEOROLOGICAL RESEARCH COMMITTEE

The use of liquids other than water with wet and dry bulb psychrometers

by

G.C. Simpson.

At the request of the Meteorological Research Committee experiments have been made at Kew Observatory to determine whether sodium per chlorate or 2:6 lutidine can be used to wet the wet bulb of a psychrometer in order to obtain reliable humidities at temperatures below the freezing point of pure water.

Two electrically driven Assmann psychrometers were obtained and used in the experiments: one was supplied with pure water and used in the usual manner, the wet bulb being emersed in water before each run; the other was supplied with the liquid under test either by wetting the bulb in the usual way or by means of a wick.

Lutidine.

Lutidine 2:4. As lutidine 2:6 was not available when the work commenced a supply of 2:4 was obtained. Very little information regarding the physical characteristics of this liquid is available. It is stated in M.R.P. 187 that lutidine 2:4 is soluble to the extent of 20 grams in 100 cc of water. This was confirmed at room temperature so far as we were able with the limited means available at Kew, and it appeared further that the solubility of lutidine in water has a maximum value in the neighbourhood of normal room temperatures as lowering and raising the temperature of a saturated solution results in precipitation.

Experiments to determine the freezing point of various mixtures of lutidine 2:4 in water, lying between 1/5 and 1/2, showed that deposition of crystals of cryohydrate became marked when the temperature was lowered to 31°F, the whole becoming solid at a temperature of $25\frac{1}{2}$°F at temperatures between 31°F. and $25\frac{1}{2}$°F. the liquid became very pasty. This high freezing point would appear alone to be sufficient to rule out lutidine 2:4 as suitable for the purpose in view.

Various solutions of the lutidine were used to wet the bulb of the psychrometer. The actual observations will not be detailed but the following gives the results of the measurements.

(a) When the bulb was wetted with pure lutidine (i.e. without added water) there was only a small lowering of the wet-bulb temperature, approximately 1°F. with atmospheric relative humidity of 64%.

(b) When water was added to the lutidine the wet bulb temperature fell to within one or two tenths of a degree F of that of the water wet bulb. This was true over the whole range of mixtures used from 1 lutidine in 1 water to 1 in 5 (saturation), and over a range of temperature from 33°F. to 48°F.

We are not clear at Kew about the physical chemistry of the results but it appears as though the lutidine is inactive when mixed with water. Alone the lutidine does not evaporate appreciably (boiling point 142°C.) and therefore undiluted does not lower the wet bulb temperature (the small lowering found may be due to the presence of a small amount of water); while a mixture carries water to the bulb which evaporates as though the lutidine were not present giving the normal wet bulb depression. The mixture therefore acts as pure water so far as the depression of the wet bulb is concerned.

Lutidine 2:6.

When these experiments were nearly completed we obtained a sample of lutidine 2:6. Rough trials showed that the freezing point of mixtures is practically the same as for lutidine 2:4; it therefore did not appear necessary to continue experiments on this substance.

Conclusions on lutidine. The addition of lutidine to water gives it no practical advantages as a solution for wetting the wet bulb of a psychrometer as the depression of the wet bulb is not significantly changed and the freezing-point of the solution is only 6½°F below that of pure water.

Sodium per chlorate.

Theoretical considerations give the vapour pressure over a saturated solution of sodium per chlorate to be .73p (p being the v.p. over pure water at the same temperature); and the freezing-point to be -34°F. The latter looked promising for the purpose in view.

A psychrometer using saturated sodium per chlorate instead of water will give relative humidities (1-.73) 100% = 27% too high, and a bulb newly wet with a saturated solution confirmed this.

When the relative humidity is above 73% water is deposited from the atmosphere on the bulb when wetted with saturated solution - this lowers the concentration of the solution and the correction to be applied cannot be determined. To try to get over this difficulty we connected the wet bulb to a reservoir of saturated solution by a thick wick (6 ordinary screen wicks) but this was unsuccessful for the difference between the two psychrometers (one with solution and one with pure water) decreased at first at the rate of about 2% in every fifteen minutes with relative humidities round about 90%. During the observations there were no humidities below 73% when the reverse action could be expected, i.e. the solution on the bulb continuously becoming more and more saturated; but from our experience we conclude that it would be equally objectionable.

We can see no means of overcoming this difficulty except by arranging a flow of solution over the bulb sufficiently copious to keep it supplied constantly with fresh saturated solution. This was tried in the case of salt solution and we evolved an apparatus which would work on the ground under supervision; but it hopelessly failed when tried in an aeroplane.

Conclusions on sodium per chlorate. Sodium per chlorate would be as effective as a salt solution for use in wetting the wet bulb of a psychrometer and have the advantage of a lower freezing point and the elimination of a thick encrustation of salt with low humidities; but the difficulty of maintaining a saturated solution on the bulb (on which the accuracy of the instrument depends) is just as great as with the salt solution. No solution of this latter problem is at present in sight.

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M.R.P. 221
23rd February 1945.

METEOROLOGICAL RESEARCH COMMITTEE

A further study of the conditions associated with ice off Newfoundland.

by G.T. Walker.

Introduction.

In an earlier note (M.R.P. 196) it was shown that arctic weather was subject to fluctuations or surges persisting over widely different lengths of time. Down to about 1905 the variations in the ice which appears off Newfoundland followed closely those of conditions in the North Atlantic, and an efficient forecasting formula was developed. Apparently a big fluctuation then developed and the present note is an attempt to throw light on the relations under the new conditions between the amount of ice and the distributions of pressure and temperature in the winter immediately preceding the ice.

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by ---------------------- date [illegible] 53

In view of the fundamental uncertainty regarding the changes of relationships with Newfoundland ice which occurred about the year 1905 it seems essential to extend the examination of correlation coefficients over a wider area, the data examined being, as before, those of the winter months preceding the outbreak of ice.

The results of such an examination will be found in Table 1, of which the first half, columns 2 to 5, gives relations with pressure and the second half, 6 to 9, with temperature. In each half the first pair of columns refer to the period ending with 1905 and the second to the period beginning with 1906. In each pair the first column gives relationships with Newfoundland ice (N.I.) and the second with the North Atlantic oscillation (N.A.O.), which represents empirically the joint effect of a number of pressures and temperatures which fluctuate together (Walker, 1932, p.54.)* coefficients based in fewer than 10 years are excluded, and those on 10 to 15 years are enclosed in brackets: the number of years lies in general between 16 and 30.

TABLE I.

Station	Pressure				Temperature			
	Before '06		After '05		Before '06		After '05	
	N.I	N.A.O.	N.I	N.A.O.	N.I.	N.A.O.	N.I.	N.A.O.
Aden					-28	-06	+12	+16
Algiers					(-18)	-22	+38	-16
Angmagsalik					(-06)	(+16)	+30	+16
Anticosti					-54	-54	-18	-46
Archangel	-38	-34	+38	-34	+48	+64	-22	+44
Astrakhan	-04	+10	+16	+24	-08	-12	+16	+10
Athens					(-54)	(-54)	-58	-40
Batavia	-18	-22	+02	-06	+26	+22	+10	+20
Bergen					+60	+82	+14	+62
Bermuda	+68	+62	-02	50	+64	+54	+24	+48
Bodo	-46	-36	+44	-28		(+48)	+14	+70
Bucharest					-12	-26	-62	-04
Buenos Ayres	-38	-30	-20	24	-14	+18	+12	+32
Bushire					-38	-38	(+22)	(-30)
Cairo	+42	+44	+02	+44	-66	-60	-20	-50
Calcutta	-28	-34	+12	-12	-06	-02	(+44)	(+18)

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* Its measure is proportional to (Vienna press)-(Stykkisholm Press)-(Ivigtut press)+(Bodo temp.) +(Stornoway temp.)+.7(Bermuda press.)-.7 (Godthaab temp.) +.7 (Hatteras + Washington temp.)/2.

	Pressure				Temperature			
	Before '06		After '05		Before '06		After '05	
Station	N.I.	N.A.O.	N.I.	N.A.O.	N.I.	N.A.O.	N.I.	N.A.O.
Cape Town	+20	-08	-06	-04	-26	-08	-18	+16
Charleston	+56	+38	-24	+42	+40	+72	+18	+44
Colombo	-18	-14	+14	+16				
Copenhagen	+42	+66	+52	+26				
Cuyaba			+12	+20				
Darwin	-16	-26	+16	00	-12	+04	+20	+16
Dawson							+28	-34
Edmonton	00	-42	-12	-12	-24	-16	00	-04
El Paso	+22	+34	-12	+28	+06	-06	-02	-04
Freetown	(-14)	(-58)	-08	+14	+14	+24	(+14)	(+52)
Georgetown	+02	-06	+12	-02				
Gjesvaer	-64	-62	+48	-28	(+42)	(+70)	+08	+60
Good Hope			(-34)	(+32)				
Grimsey					+30	+56	+30	+32
Habana	+48	+34	-06	+04				
Halifax	+28	+54	-04	+56				
Hebron (Lab.)					-66	-84		
Helsinki	-10	+08	+44	-08	+62	+66	-22	+50
Honolulu	+28	+06	+06	-08	-36	-08	+02	+14
Irkutsk	-48	-42	+06	-48	+58	+54	-14	+14
Ivigtut	-82	-76	-14	-56	-56	-76	+08	-38
Jacobshavn	-80	-70	-14	-48	-58	-78	+42	-02
Kasan	-32	-20	+06	-28	+32	+30	-26	+30
Lahore	-28	-34	+18	-10				
Madeira					-04	-08	(-04)	(-04)
Madrid	+76	+80	-08	+50	+22	-16	(-36)	(-50)
Manila	-04	-16	+34	-10	-06	+16	-02	+22
Mauritius	-36	-20	+08	-06	(00)	(-02)	-02	-16
Mexico					-06	+10	-08	-12
Moose Factory	(00)	(00)	-12	+20	(-16)	(-18)	-10	+06
Moscow					+44	+38	-18	+48
Nome			-28	+08			-42	+18
Obdorsk	-56	-52	(-20)	(-14)	+48	+50	+06	(-48)
Paris					+38	+08	-12	+32
Perth(W.Austr.)	+08	-06	+16	00	+10	+18	+30	+04
Petropavlovsk			(-34)	(+46)	(-14)	(-14)	(-08)	(-22)
P.Delgada	+38	+50	-30	+30				
Punta Arenas	00	-34	-16	+08				
Quixeramobim			+12	+34				
Rio de Janeiro	-34	-30	-36	-04	+58	+26	+44	+02
St. Helena	(-28)	(-10)	-12	+10	(+06)	(+08)	+10	+04
St. John's(Newf.)					-66	-66	-32	-48
St. Louis					+30	+66	00	+42
St. Vincent					-34	-40	-10	-58?
Samoa	(-10)	(-16)	-24	+02	00	+06	+40	+28
San Fransisco					-10	-54	00	00
Santiago	+12	-12	-20	-24	-26	-12	+04	+10
Sitka					-16	-40	+12	-32
Spitzbergen			+08	-64			-24	+46
Stykkisholm	-76	-88	00	-72				
Sydney (N.S.W.)	-14	-08	+02	+14				
Tashkent					-12	-18	(-16)	(-08)
Thorshavn	-42	-36	+36	-32	+62	+88	+32	+46
Tobolsk					(+14)	(+22)	-26	-02
Tokyo	+04	+10	+30	+56	+48	+48	(+02)	(-12)
Tomsk	-54	-40	+02	-28	+46	+36	-22	+18
Trinidad					-26	-32	(-22)	(+02)
Upernivik	-80	-76	-08	-54	-52	-76	+32	-14
Valentia	+14	+36	+20	+02	+68	+66	+22	+62
Verkhoyansk			(+44)	(+84)	+02	-02	(+22)	(-18)
Vienna	+72	+86	+40	+54	+16	+04	-46	+18
Vladivostok					+42	-12	+20	-40
Wellington	+10	+06	-14	00	-04		-52	
Winnipeg	00	-24	-02	-10	-16	+06	-02	+10
Yakutsk	-70	-66	-20	-02	+60	+50	-04	-14
Zanzibar	-08	(-10)	-04	00	(+42)	(+44)	+28	(+52)
Zikawei	-30	00	+08	+10	+14	+40	+36	+22

2. The eight columns of Table 1 are plotted in Figures 1 to 8, and it will be seen that, in accordance with what has previously been noted, Figs. 1,2 and 4 have much in common, and Figs. 5, 6 and 8; but Figs. 3 and 7 stand alone. In other words the break of 1905 affects relations with N.I., but not with the N.A.O., and the close similarity of the relationships with N.I. and the N.A.O. applies only to the first period. Accordingly we find between the variations of N.I. and N.A.O. a correlation coefficient of .82 before the break and only .16 after it.

We must not however dismiss as a mere statistical paradox the change which occurred about 1905, a change to which I know of no parallel. Between Charts 1 and 3 the differences in the coefficients are as much as .7 in Greenland, .6 near Bermuda, .8 in Spain and .9 in Norway. Between Charts 5 and 7 the differences, though not so striking, are really big, amounting to .8 in Greenland, and .6 in central Europe and regions in Siberia. See §7 below.

The charts (Figs. 2, 6) illustrating relations with the N.A.O. show, as Wiese urged, clear relationships with pressure and temperature in Siberia; but Exner's suggested zonal temperature relationships - roughly with latitudes 30° and 60°N. - are not substantiated by Fig.6, there being marked negative as well as positive c.c's in both belts.

As the probable error of a coefficient is something like .15 (and of those in brackets .19) the c.c's in areas south of the equator are not in general big enough to have much meaning.

3. The similarity between Figs. 1 and 2, and between Figs. 5 and 6, suggests a further question. When K, the c.c. between N.I. and the N.A.O., has a high value such as .82, it is clear that if there is a direct control of a pressure by the N.A.O. this will produce variations in that pressure which will be closely correlated with N.I. and will wrongly suggest control of the pressure by the N.I. Thus we must attempt to disentangle the effects.

If the direct effect of N.I. is to produce a partial coefficient $\underline{r}$ between its variations and those of the pressure, and the direct effect of the N.A.O. a partial coefficient $\underline{r}^1$ then the total coefficients $\underline{c}$, $\underline{c}^1$ between the pressure and N.I, and between it and the N.A.O. (those tabulated in Table I) will be

$$c = r + Kr^1, \quad c^1 = Kr + r^1$$

so that in the earlier period, when $k = .82$, with sufficient accuracy,

$$r = 3c - 2.5c^1, \quad r^1 = -2.5c + 3c^1 \ldots\ldots \qquad (1)$$

For the later period, when $k = .16$, we find similarly

$$r = c - .15c^1, \quad r^1 = -.15c + c^1 \qquad (2)$$

4. It may be well to consider what errors in $\underline{r}$, $\underline{r}^1$ are likely. For the later period they will be little larger than those in $\underline{c}$, c^1; but for the earlier period equation (1) shows that they will be appreciably larger.

The errors in the series of values of the N.A.O., each term being the mean of a number of departures, are likely to be small; and the error in a large coefficient like .82 is not likely to be large. But if $K = .78$, $\underline{r}$ in (1) becomes $2.5c - 2.0c^1$, differing from its former value by as much as $.5 (c - c^1)$. The general impression left at this stage is that owing to inaccuracies from all sources the errors in $\underline{r}$, $\underline{r}^1$ in the earlier period are likely to be two or three times greater than those of $\underline{c}$, $\underline{c}^1$. This subject will be revived in § 6(a) below.

5. In order to increase reliability we shall now work as far as possible with fairly homogeneous groups of stations representative of characteristic

/areas.

areas. For each of the areas in the first columns of Tables 2, 3 the stations are given in the second columns, and the means for the area of the values of c, c^1 have been computed. Then equations (1), (2) give for the earlier and later periods the values of r, r^1 which are inserted (to one decimal place) in the last four columns of the tables.

Table 2. Partial coefficients with pressure.

Area	Stations	Earlier		Later	
		N.I	N.A.O.	N.I	N.A.O.
U.S.S.R., Asia	Irkutsk, Tomsk	-.5	.0	.1	-.4
High-pressure-belt	Bermuda, Madrid, Vienna	.3	.5*	.0	.5
N.W. Europe	Copenhagen, Bodo, Gjesvaer	-.4	.3	.5	-.2
Iceland, Faroes	Stykkisholm, Thorshavn	-.2	-.4*	.3	-.5
Greenland	Ivigtut, Jacobshavn, Upernivik	-.6	-.2*	.0	-.5
S.E. Canada	Halifax (N.S.)	-.5	.9	-.1	+.6
Central Canada	Edmonton, Winnipeg	.8	-.9	-.1	-.1
S.E. of U.S.A.	Charleston	.7	-.3	-.3	+.5
Alaska	Nome			-.3	+.1

Table 3. Partial coefficients with temperature.

Area	Stations	Earlier		Later	
		N.I	N.A.O.	N.I	N.A.O.
U.S.S.R., Asia	Irkuksk, Tomsk	.4	.0	-.2	.2
Central Europe	Paris, Vienna, Bucharest	.5	-.5	-.4	.2
N.W. Europe	Valentia, Bergen, Gjesvaer, Helsinki	0	.7*	0	.6
Iceland, Faroes	Grimsey, Thorshavn	-.3	.8	.2	.3
Greenland	Ivigtut, Jacobshavn, Upernivik	.3	-.9*	.3	-.2
Labrador	Hebron	.1	-.9		
S.E.Canada	St.John's (Newf.), Anticosti	-.3	-.3	-.2	-.4
S.E., U.S.A.	St. Louis, Charleston	-.5	1.0*	0	.4
Alaska	Sitka	.5	-.8	.2	-.3
	Nome			-.4	+.2

6. When considering Tables 2,3 we shall discuss first the normal situation, of which the earlier period is typical.

(a) First regarding the liability to error, we can best get light on this by examining the coefficients of pressure and temperature with the N.A.O. of which the chief relationships are well known: they are those with high pressure from Bermuda to Vienna, low pressure in Iceland and Greenland, high temperature in N.W. Europe and the S.E. of the U.S.A., and low temperature in Greenland. The corresponding coefficients are marked with asterisks in the Tables, and all have the signs expected; those with temperature are particularly well marked and the weakest agreement is that of -.2 with low Greenland pressure. Thus in spite of the arguments at the end of §4 it would seem that the error to be expected in a coefficient will not seriously exceed .3.

(b) The relations with Newfoundland ice may now receive attention. Here a conspicuous feature is the group of pressure coefficients, -.6 with Greenland (and -.5 with Halifax, N.S.) combined with +.8 with central Canada (and .7 with the southeast of the U.S.A.). It has generally been believed that abundant ice off Newfoundland was associated with strong northerly winds between Greenland and Labrador and a strong pressure gradient; the favouring low pressure in Greenland is conspicuous in Fig. 1, but there is no corresponding region of marked high pressure. As Table 2 shows, there is a very strong negative relationship (with a partial coefficient -.9) between the N.A.O. and central Canadian pressure; and, in view of the c.c. of .82, it seems that this has hitherto obscured the positive coefficient of .8 between Newfoundland ice and central Canadian pressure.

7. If the situation before '06 was produced by a piece of physical machinery and the subsequent results show a fundamental alteration in character, we must obviously attribute this, not to a breakdown of statistical laws but to some essential change in the mechanism in operation. Now it appears to be widely recognised that the provision of the ice off Newfoundland is a complicated matter. Thus E.H. Smith (1928, 39) definitely concludes that "pack ice from the east Greenland current never crosses to American waters", and maintains that as sources of ice we have not only the current from the north polar cap flowing southward down Smith Sound, but also strong currents from the west flowing through the islands of the Canadian archipelago and emerging along Jones Sound, Lancaster Sound and Hudson Strait.

There may also be in existence important fluctuations such as the material rise in temperature reported in the Greenland Sea and the ocean north of the U.S.S.R.; this might diminish materially the ice-producing effect of the polar current down Smith Sound, previously controlled merely by pressure conditions. Another source of anomaly might be variations in the eastward flow of the ocean along the north coast of Canada and into the archipelago. Unfortunately the scarcity of meteorological records makes a satisfactory investigation of such fluctuations by statistical methods very difficult.

Appendix on the method of computing correlation coefficients.

In previous work on seasonal relationships, when the number n of years under examination, roughly 50, was nearly the same for all variables, it sufficed to choose the unit of measurement such as to make the standard deviation, σ, equal to $\sqrt{20}$: the formula used was then simply, c.c. = $\Sigma xy/.20n$. In the present case however the values of σ for the earlier and later portions of the same series often differ widely; so the actual variances have always been computed.

When forming tables of partial coefficients, such as 2, 3 above, there is a temptation to combine available correlation coefficients from several stations though the years on which they are based are not identical, as they theoretically should be. This lack of consistency may show itself in the appearance of values of r, r^1, exceeding unity.

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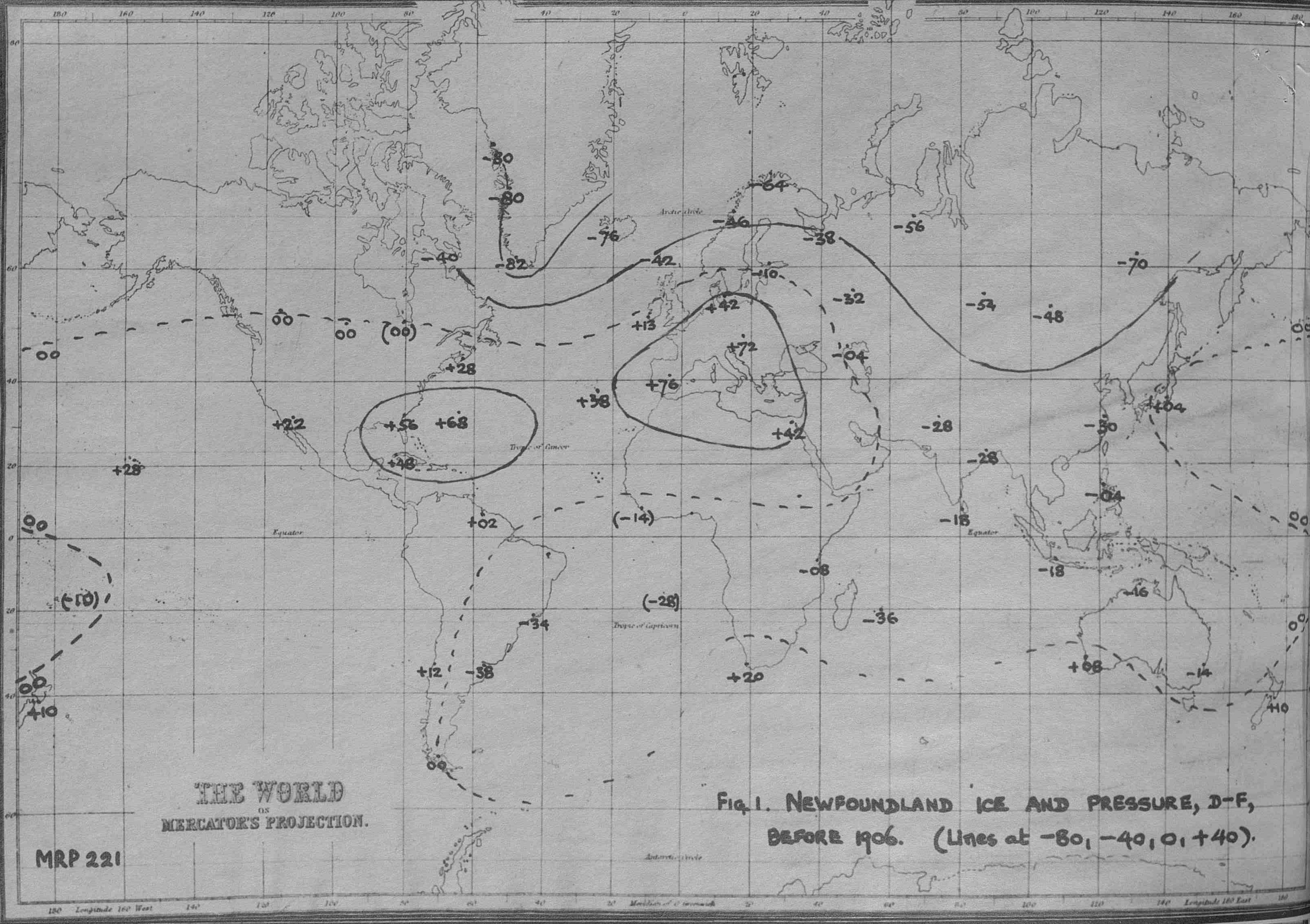
THE WORLD ON MERCATOR'S PROJECTION.
Arctic Circle
Tropic of Cancer
Equator
Tropic of Capricorn
Antarctic Circle
Meridian of Greenwich
Longitude 160 West
Longitude 160 East
-80
-80
-82
-40
-76
-42
-46
-64
-38
-56
-70
-10
-32
-54
-48
00
00
00
(00)
00
+13
+42
+72
-04
+28
+76
+38
+22
+56
+68
+42
+48
+28
-28
-30
+04
-28
-04
-18
(-14)
+02
00
(-10)
-08
-18
-16
-28
-34
-36
00
+12
-38
+20
+08
-14
00
+10
+10
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Fig. 1. NEWFOUNDLAND ICE AND PRESSURE, D-F, BEFORE 1906. (Lines at -80, -40, 0, +40).

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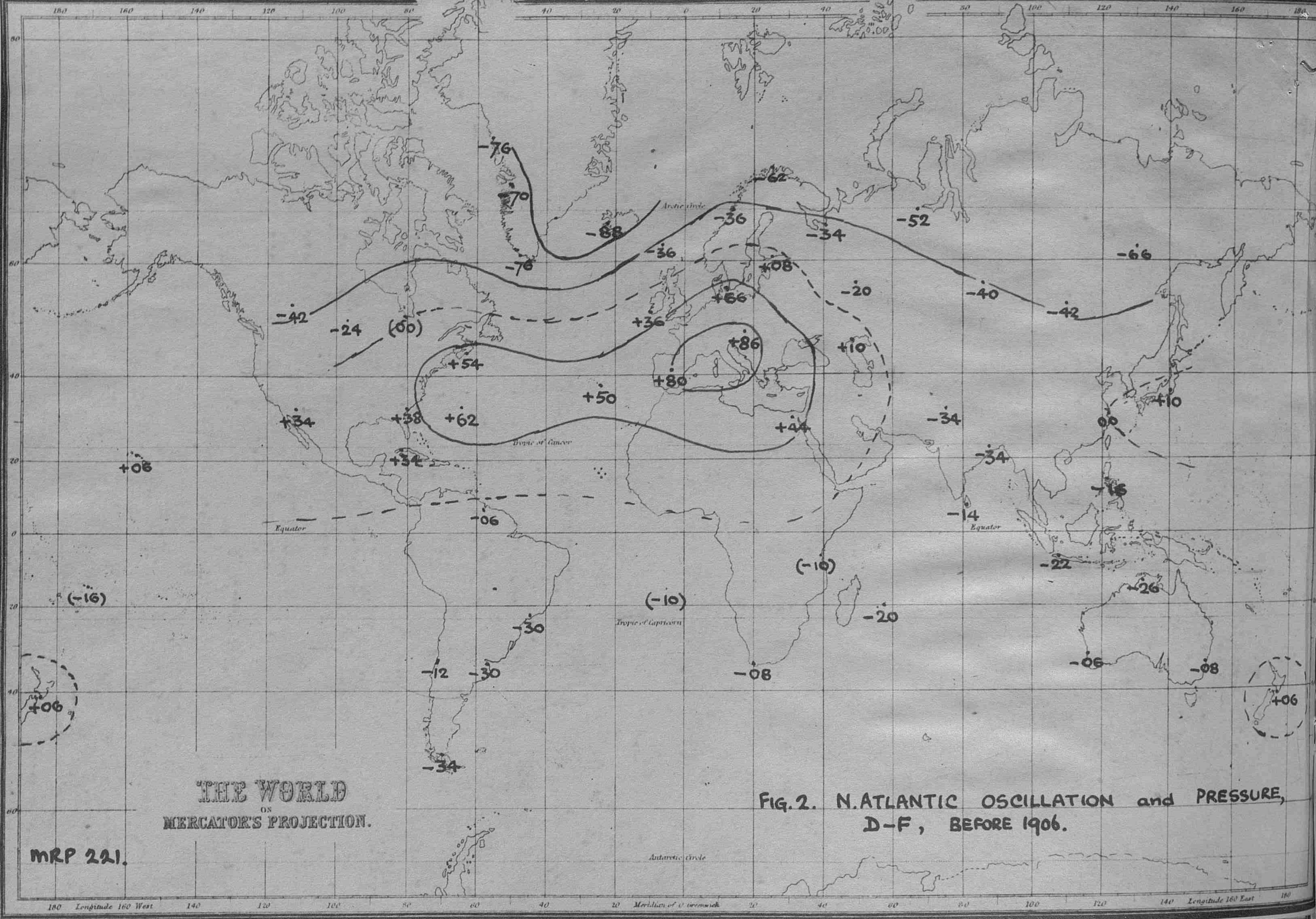
THE WORLD
ON
MERCATOR'S PROJECTION.
Arctic Circle
Tropic of Cancer
Equator
Tropic of Capricorn
Antarctic Circle
Meridian of Greenwich
Longitude 160 West
Longitude 160 East
-76
-70
-88
-62
-36
-34
-52
-66
-76
-36
+08
-20
-40
-42
-42
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+50
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+34
+38
+62
+44
-34
00
+08
+34
-34
-16
-06
-14
-22
(-10)
+26
(-16)
(-10)
-20
-30
-06
-08
-12
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-08
+06
+06
-34

FIG. 2. N. ATLANTIC OSCILLATION and PRESSURE, D-F, BEFORE 1906.

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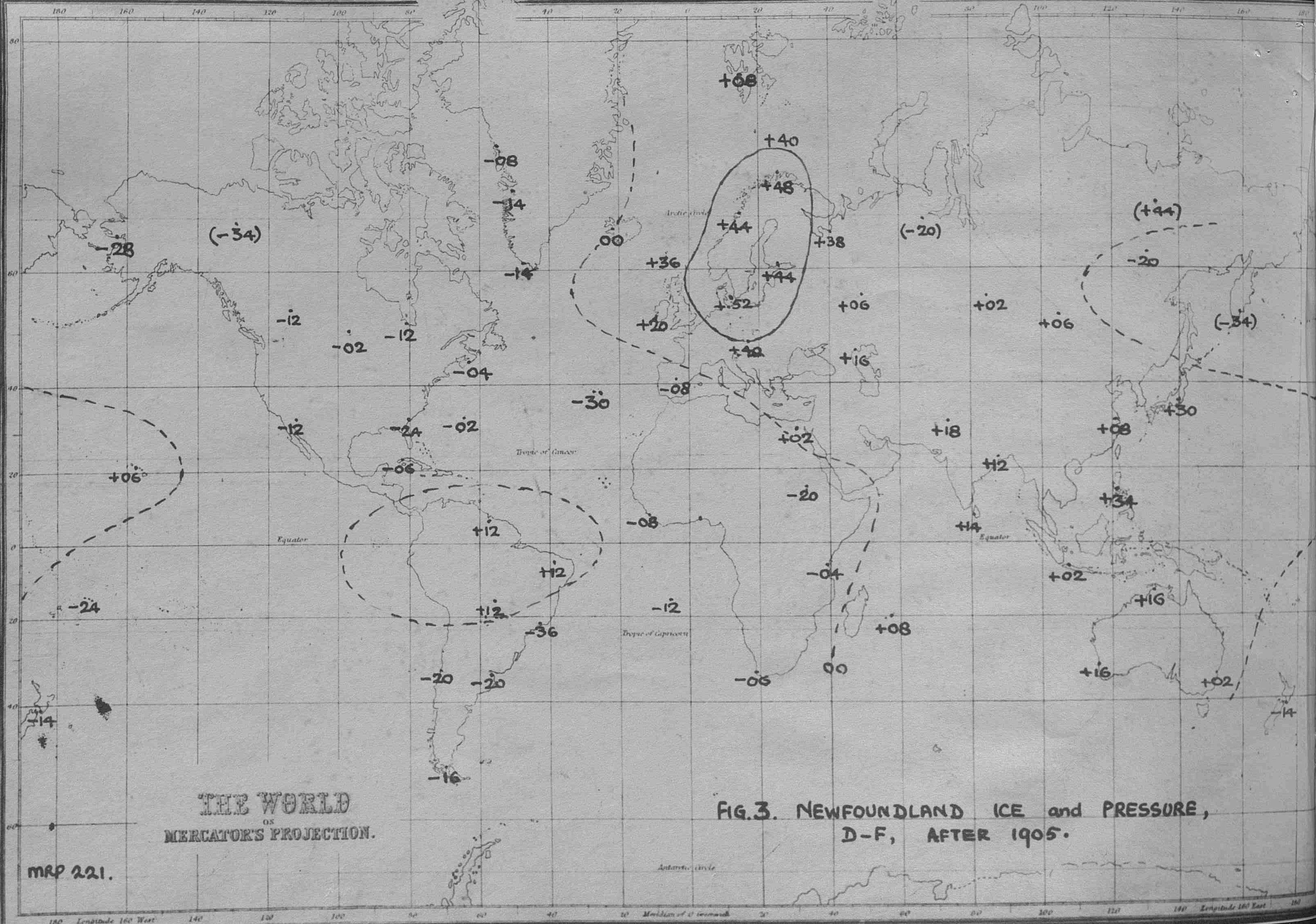
THE WORLD ON MERCATOR'S PROJECTION.
Arctic Circle
Tropic of Cancer
Equator
Tropic of Capricorn
Antarctic Circle
Longitude 160 West
Meridian of Greenwich
Longitude 160 East
+08
+40
+48
+44
+44
+52
+42
+38
+36
+20
-08
(-20)
(+44)
-20
(-34)
-08
-14
-14
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-02
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+06
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+14
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-12
-20
+02
-04
-06
00
+08
+06
+16
+02
+18
+12
+14
+06
+08
+30
+34
+02
+16
+16
+02
-14

FIG. 3. NEWFOUNDLAND ICE and PRESSURE, D-F, AFTER 1905.

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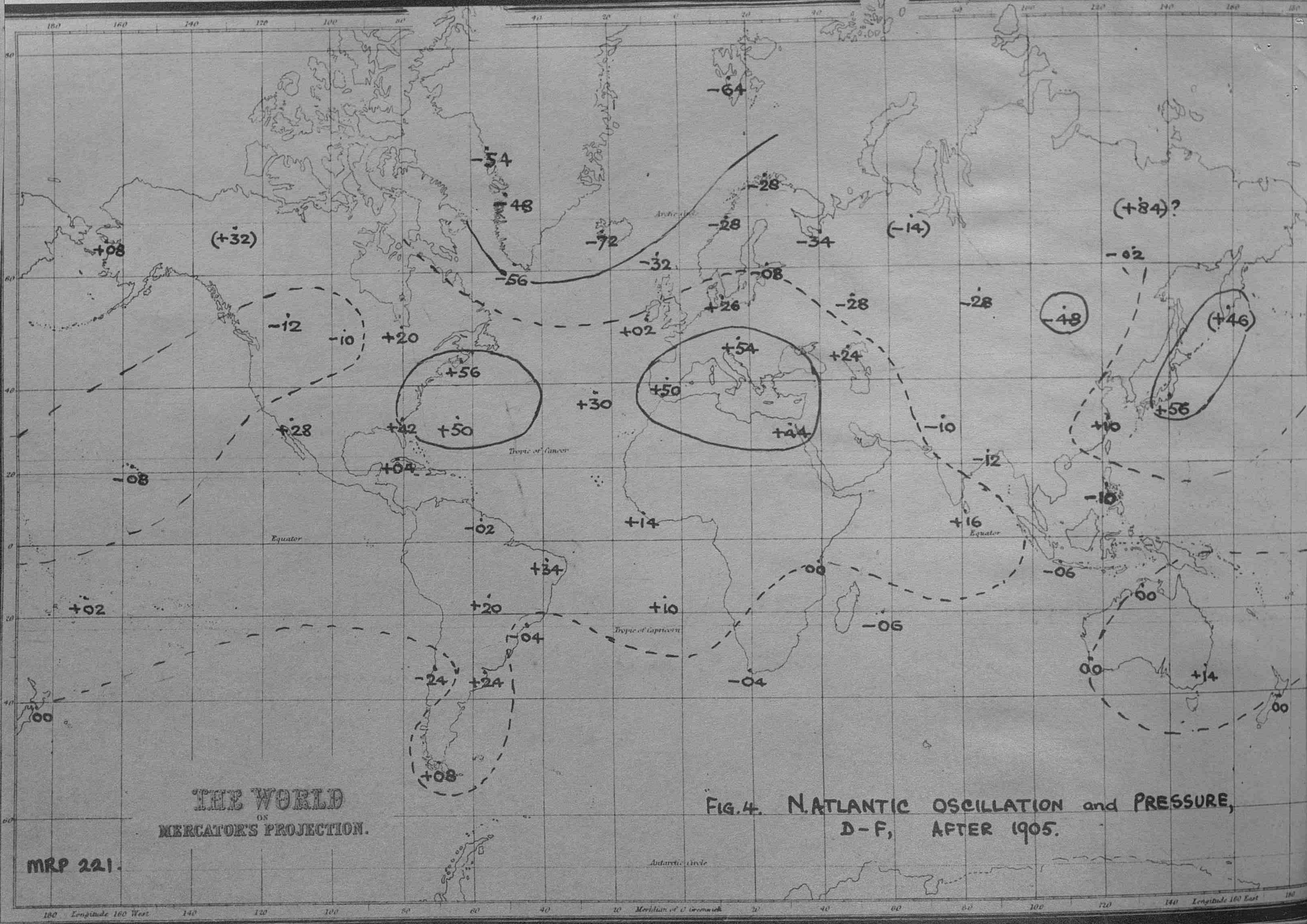
THE WORLD ON MERCATOR'S PROJECTION.
MRP 221.
Arctic Circle
Tropic of Cancer
Equator
Tropic of Capricorn
Antarctic Circle
Meridian of Greenwich
Longitude 160 West
Longitude 160 East
-64
-54
48
-28
-28
-34
(-14)
(+84)?
(+32)
+08
-72
-32
-08
-02
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-48
(+46)
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+20
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+10
+56
+28
+42
+50
-12
-08
+04
-10
+14
+16
-02
-06
+34
00
00
+02
+20
+10
-06
-04
-04
00
+14
-24
+24
00
00
+08

FIG. 4. N. ATLANTIC OSCILLATION and PRESSURE, D-F, AFTER 1905.

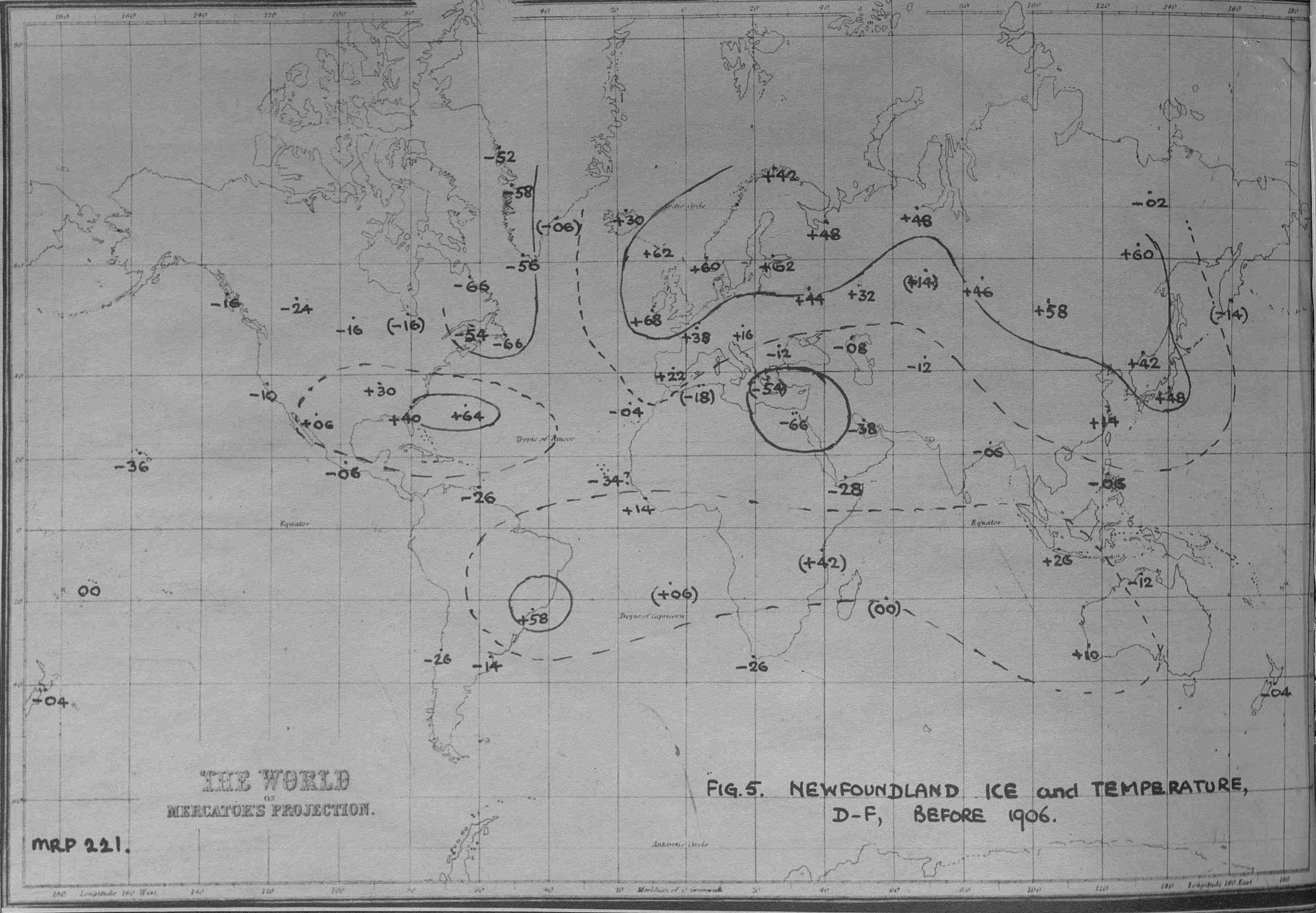
THE WORLD
ON
MERCATOR'S PROJECTION.
Arctic Circle
Tropic of Cancer
Equator
Tropic of Capricorn
Antarctic Circle
Meridian of Greenwich
Longitude 160 West
Longitude 160 East
-52
58
-02
+42
+48
(-06)
+30
+48
+62
+60
+62
+60
-56
+44
+32
(+14)
+46
-66
+58
-16
-24
-16
(-16)
+68
+38
+16
(-14)
-54
-66
-12
-08
-12
+42
+22
(-18)
-54
+48
-10
+30
+64
+06
+40
-04
-66
-38
+14
-36
-06
-06
-34?
-28
-08
-26
+14
+26
(+42)
00
(+06)
-12
+58
(00)
+10
-26
-14
-26
-04
-04

FIG. 5. NEWFOUNDLAND ICE and TEMPERATURE,
D-F, BEFORE 1906.

MRP 221.

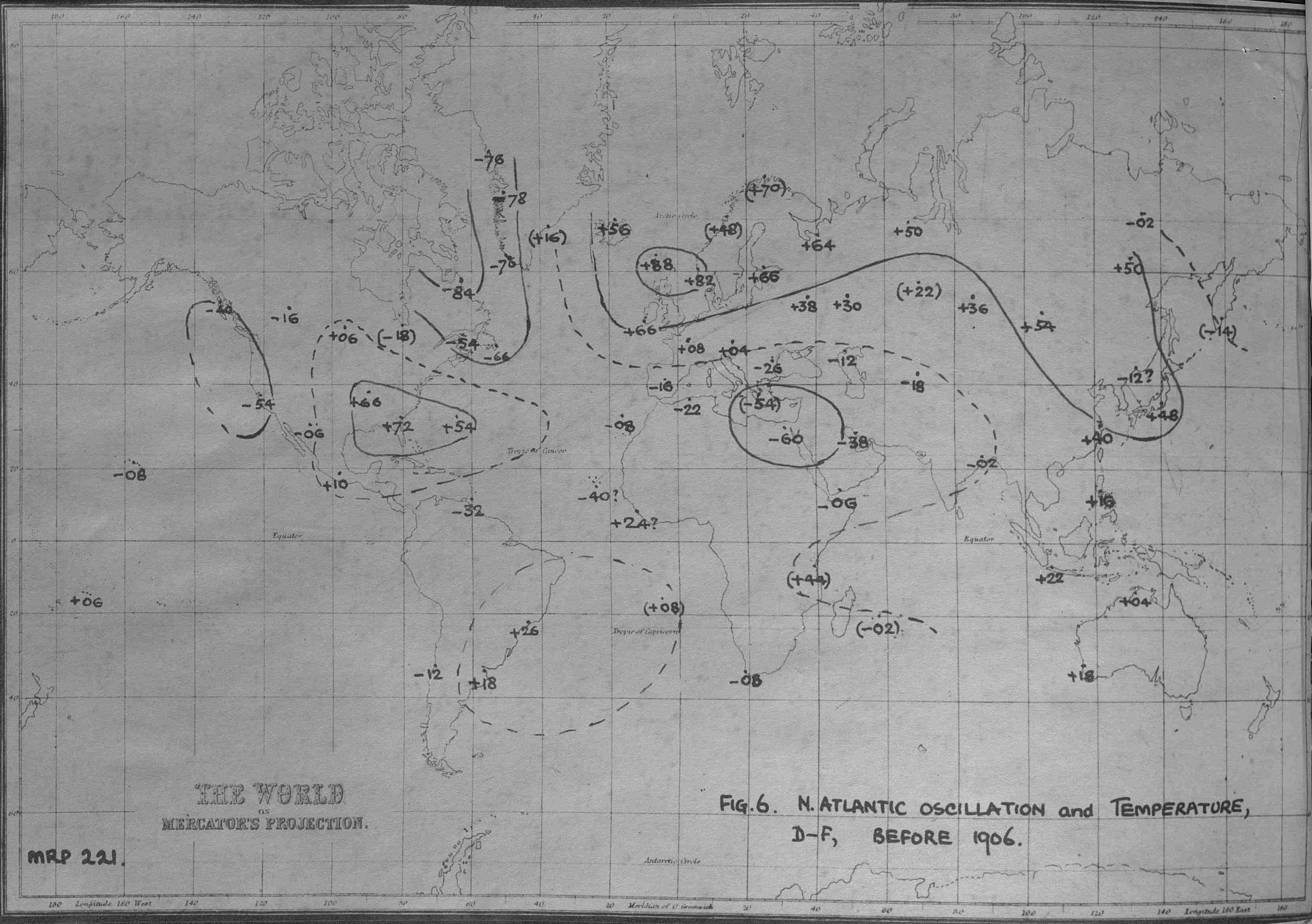
THE WORLD
ON
MERCATOR'S PROJECTION.
Arctic Circle
Tropic of Cancer
Equator
Tropic of Capricorn
Antarctic Circle
Meridian of 0 Greenwich
Longitude 160 West
Longitude 160 East
-76
-78
(+16)
+56
(+70)
(+48)
+64
+50
-02
+50
-16
-48
+06
(-18)
-84
-78
+88
+82
+66
(+22)
+38
+30
+36
+54
(-14)
-54
-66
+66
+08
+04
-26
-12
-18
-12?
-54
+66
+72
+54
-06
-08
-16
-22
(-54)
-60
-38
+48
+40
-08
+10
-32
-40?
+24?
-06
-02
+16
+06
(+08)
(+44)
(-02)
+22
+04
+26
-12
+18
-08
+18

FIG. 6. N. ATLANTIC OSCILLATION and TEMPERATURE, D–F, BEFORE 1906.

MRP 221.

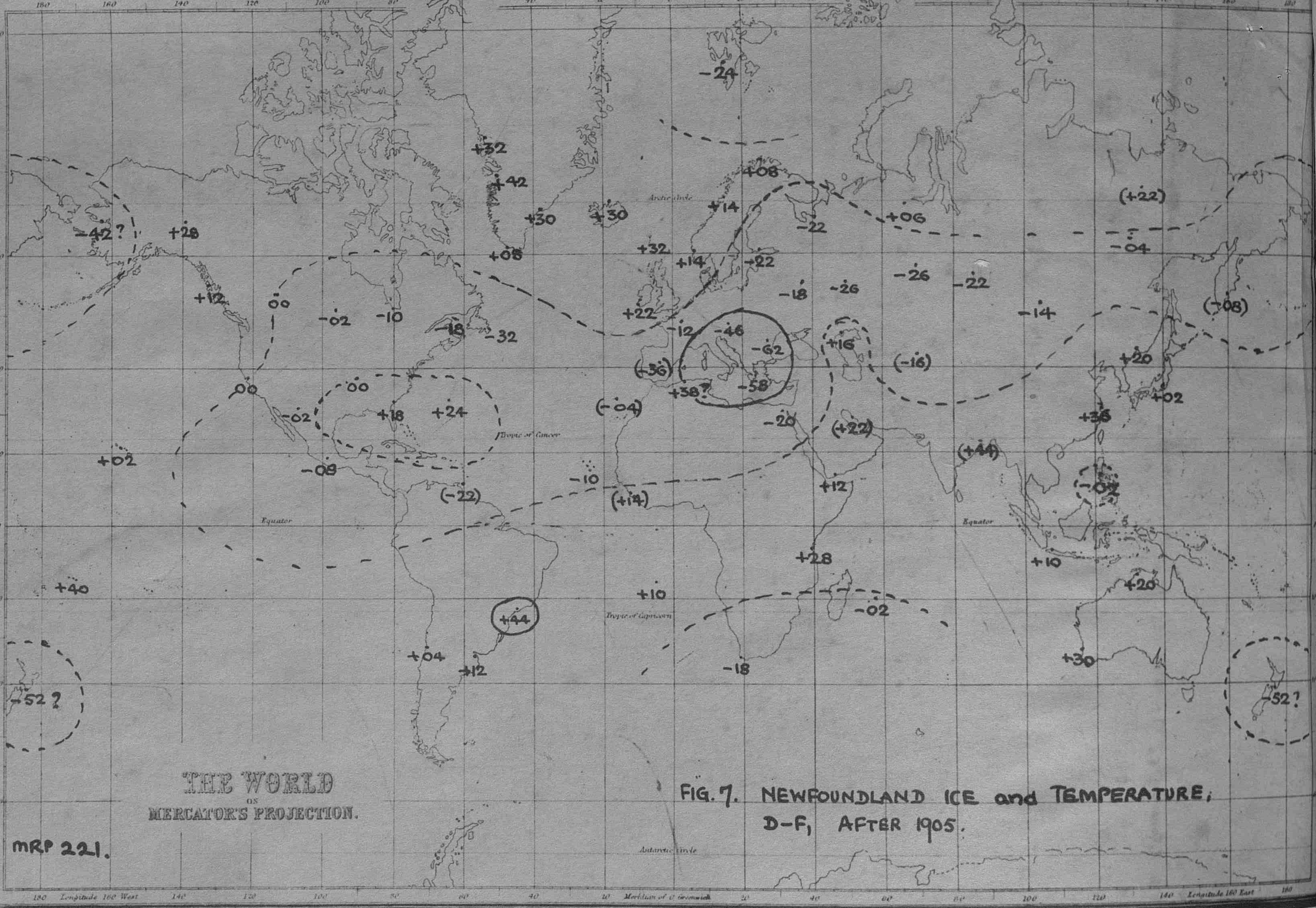
THE WORLD
ON
MERCATOR'S PROJECTION.
Arctic Circle
Tropic of Cancer
Equator
Equator
Tropic of Capricorn
Antarctic Circle
Longitude 160 West
Meridian of Greenwich
Longitude 160 East
-·24
+·32
+·08
+·42
(+·22)
-·42?
+·28
+·30
+·30
+·14
-·22
+·06
-·04
+·32
+·14
+·22
+·08
-·26
-·22
+·12
·00
-·02
-·10
-·18
-·26
+·22
-·18
-·32
·12
-·46
-·14
(-·08)
-·62
+·16
(-·16)
(-·36)
·00
·00
+·38?
-·58
+·20
-·02
+·18
+·24
(-·04)
-·20
(+·22)
+·38
+·02
(+·44)
+·02
-·08
-·10
+·12
(-·02)
(-·22)
(+·14)
+·28
+·10
+·40
+·10
+·20
-·02
+·44
+·04
+·12
-·18
+·30
-·52?
-·52?

FIG. 7. NEWFOUNDLAND ICE and TEMPERATURE; D-F, AFTER 1905.

MRP 221.

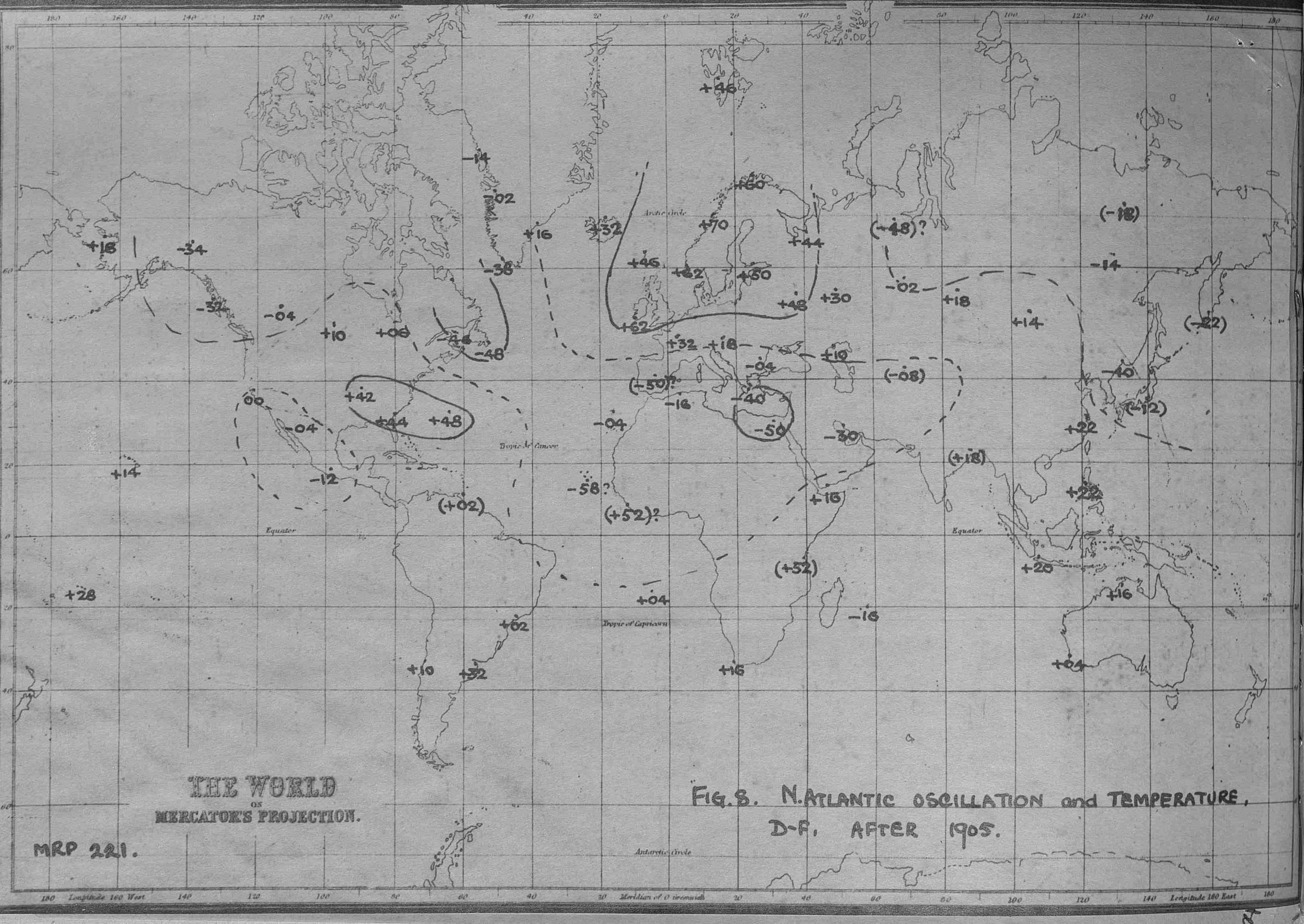
THE WORLD ON MERCATOR'S PROJECTION.
Arctic Circle
Tropic of Cancer
Equator
Tropic of Capricorn
Antarctic Circle
Meridian of 0 Greenwich
Longitude 160 West
Longitude 160 East
+18
-34
-32
-04
+10
+08
-14
·02
+16
-38
-48
-46
+42
+44
+48
00
-04
-12
+14
+28
(+02)
+62
+10
+32
+32
+46
+60
+70
+46
+62
+50
+44
+48
+30
+62
+32
+18
-04
+10
(-50)?
-16
-04
-40
-50
-30
-58?
(+52)?
+04
+16
(+52)
-16
+16
(-48)?
-02
+18
(-08)
+14
(+18)
(-18)
-14
(-22)
-40
(-12)
+22
+22
+20
+16
+04

Fig. 8. N. Atlantic Oscillation and Temperature, D-F, after 1905.

MRP 221.

** The "success" is that

Downgraded to unclassified per letter from Met. O. the nov 5 signed by P.S. Chesman

~~[illegible] MRP 305~~

ACCURACY OF UPPER AIR FORECASTS.

I. Explanatory Note.
Tables I - III. Tables of Accuracy of Forecasts for Larkhill, August - December, 1944.
II. Discussion of the Results for August - December, 1944.

I. EXPLANATORY NOTE.

Reports on the accuracy of upper air forecasts for June and July were prepared in August, 1944. These reports, which referred to the levels 750, 500 and 300 mb. above Larkhill, contained the frequency of the differences between forecast and observed values of temperature, heights and wind direction and speed, and also the 50% errors in the forecast values.

As from August, 1944, the form of the reports on the accuracy of upper air forecasts has been extended. These reports have been submitted at the end of each month, and the accuracy has been expressed in terms of the statistical characteristics set out below.

1. The forecasts are evaluated from the Upper Air prognostic charts (PRONTOURS) which are issued four times daily and apply to 12 hours after the times to which the observations refer. The pressure surfaces and the approximate (I.C.A.N.) heights to which the forecasts apply are :

Pressure Surface (mb.)[*]	I.C.A.N. Height (Ft.)
750	8086
500	18278
300	30049

The forecast values of the heights of pressure surfaces, the temperature and the wind direction and speed are noted for Larkhill and compared with the observations at the times to which the forecasts refer. The results for August to December are given in Tables I, II and III. In the current monthly reports no allowance is made for observational errors.[†] The errors indicated represent, therefore, the sum of the forecast and observational errors.

2. The statistical characteristics used are :

(1) Percentage frequency of deviation of forecast from observed values.
(2) Probable error of forecasts.
(3) Standard error of forecasts.
(4) 12-hour variance.
(5) Success[**] of forecasts.
(6) Percentage success of forecasts.

In view of the lesser importance attached to the temperature forecasts, and in order to reduce the amount of computations, only the characteristics (1) and (2) will be evaluated for the temperature forecasts.

3. Definitions. Let v denote the 12-hour variation of the observed element, e the observational error, and N the number of observations. Then

$$E = \sqrt{\Sigma e^2/N} \quad = \text{standard error.}$$

$$P = 0{\cdot}6745\,E \quad = \text{probable error.}$$

$$V = \sqrt{\Sigma v^2/N} \quad = \text{12-hour variance.}$$

---

[*] From 1st January 1945 the 750 mb. chart will be replaced by a chart for the 700 mb. surface, the I.C.A.N. height of which is 9876 ft.

[†] The observational errors form the subject of a separate report.

[**] The "success" is that part of the variance of the observation that is correctly forecast, assuming that there are no errors in the observation. Thus, if the forecasts were perfect and if there were no observational errors, the "success" would be equal to the variance, and the "percentage success" would be 100.

The apparent variance is composed of the true variance (V_T) plus a component due to the errors, each variation figure containing two individual observations and therefore also two individual errors.

Let F denote the standard error of the forecasts. The standard forecast error is composed of

(a) the true forecast error (F_T).

(b) a component due to the errors in the observations with which the forecasts are compared, and

(c) a component due to the errors in the analyses on which the forecasts are based.

Assuming a normal distribution of the errors, the following relations hold :

$$V^2 = V_T^2 + 2E^2$$

and

$$F^2 = F_T^2 + E^2(1 + k)$$

where k denotes an adjustment factor representing the quality of the analyses. If the analyses adhere strictly to the observations, the standard error of the analyses is the same as that of the observations, and

$$k = 1.$$

On the other hand, if the observational errors were completely removed through the process of analyses

$$k = 0.$$

Owing to the theoretical relation that exists between the wind and the slope of pressure surfaces, and the fact that this relation is used for fitting and deriving the winds, the analyses must represent a compromise solution, and it is not in general possible to reduce k to zero for both elements. The more the lines are adjusted to reduce k for one element, the larger may k be for the other element.

In a fairly dense network of stations, the observational errors are largely minimised through the process of analysis, and

$$1 > k > 0.$$

If, however, observations are sparse (e.g. over the Atlantic Ocean), observational errors cannot be reduced appreciably through synoptic smoothing, and, in addition large errors of analyses may be introduced through interpolating between widely scattered points of observations. In such cases

$$k > 1.$$

†Each figure for the variation can be considered as consisting of two components

$$v = e + s$$

---

† The analysis of variance in this paragraph strictly applies only to scalar quantities. The variance of a vector quantity requires slightly more complex analysis but in this application, in the absence of any systematic error such as consistent under-estimation of wind change, the result for variance of the wind vector is substantially the same as for a scalar.

where s is that part of the change which is successfully foreseen by the forecaster. Hence

$$V_T^2 = F_T^2 + S_T^2 + 2 r F_T S_T$$

S_T being the variance of s, and r the correlation coefficient between s and e. r is positive or negative according as the variations are consistently under- or over-estimated. In the absence of any such consistency

$$V_T^2 = F_T^2 + S_T^2$$

whence, in terms of the observable quantities V and F

$$S_T^2 = V^2 - F^2 - E^2 (1 - k).$$

For want of a better term, S_T will be called the true success of the forecast. The limitations of this concept are apparent from the above derivation, but it is desirable to introduce a statistic of this type as a basis of comparison from month to month, and in order to provide a yardstick by which progress can be judged.

The forecasts for Larkhill will frequently depend upon the analyses to the west (i.e. to the windward) of the British Isles where observations are so sparse that the errors in the analyses are larger than the observational errors, i.e. $k > 1$. Since we cannot assess the average magnitude of k, we shall put $k = 1$, giving

$$S = \sqrt{V^2 - F^2}$$

as the computed success, which will in most cases be smaller than the true success.

The success should be compared with the variance of the element concerned. The "Percentage Success" is defined as 100 times the ratio of the success to the variance, viz.,

$$\text{Percentage Success} = 100 \, S/V = 100 \sqrt{1 - \frac{F^2}{V^2}}$$

It should be noted that, in practice, very large observational errors occur more frequently than indicated by the normal distribution. Such errors may be assumed to contribute equally to V, F and E, and thus since

$$S_T^2 = S^2 - E^2 (1 - k)$$

the computed success when $k > 1$ is still further diminished in relation to the true success. Even when $k < 1$, the computed ratio S/V is usually smaller than the true ratio of success to variance, so that these abnormal errors have a systematic bias in the sense of an adverse reflection on the success of forecasting.

NOTE : In view of the fact that the upper air network has its maximum density over the British Isles while the circulation of the atmosphere is predominantly from the west, it is to be expected that the accuracy of the forecasts is greatest along the East Coast of the British Isles, over the western part of the North Sea, and also to the South and Southeast of England. In these regions the accuracy would normally be higher than that which refers to Larkhill. An accuracy lower than this would normally be found to the West and North of the British Isles and in other regions where observations are sparse.

C H B Priestley

M.O.12.
23rd February, 1945.

TABLE I - Frequency of Errors in Forecast Heights of Pressure Surfaces.

Height (Feet)	AUGUST			SEPTEMBER			OCTOBER			NOVEMBER			DECEMBER		
	Frequency %			Frequency %			Frequency %			Frequency %			Frequency %		
	750	500	300	750	500	300	750	500	300	750	500	300	750	500	300
0-20	37	21	10	12	9	7	27	21	9	19	14	7	31	22	8
21-40	20	12	14	25	5	9	18	3	9	12	10	5	15	15	7
41-60	6	13	2	12	12	7	12	12	5	16	11	4	13	7	7
61-80	13	10	3	17	9	7	8	12	5	14	10	10	18	9	12
81-100	9	6	10	12	17	7	11	14	8	11	6	8	7	13	8
101-120	8	10	5	12	12	5	5	3	1	7	5	5	5	8	8
121-140	3	6	9	3	7	4	12	9	8	5	11	10	3	7	8
141-160	1	7	10	-	2	9	1	3	5	3	8	4	1	5	10
161-180	1	6	9	2	5	4	3	6	1	-	6	8	1	3	9
181-200	1	1	7	-	5	-	-	2	11	3	3	7	1	4	4
201-220	-	1	3	3	5	4	-	2	5	3	3	7	-	1	-
221-240	1	3	2	-	2	4	-	-	3	-	1	-	2	-	1
241-260	-	3	3	-	3	5	-	3	3	-	1	1	2	1	2
261-280	-	-	2	-	2	4	3	2	1	3	-	1	1	2	3
281-300	-	-	1	-	2	4	-	2	1	-	1	3	-	-	4
>300	-	1	10	2	3	20	-	6	25	4	10	20	-	3	9
Probable Error	49	79	124	64	101	160	61	111	192	90	116	179	57	80	121
Standard Error	73	117	184	95	150	237	90	165	285	133	172	265	85	118	179
12-hour Variance	80	131	218	126	187	304	162	285	408	157	239	353	149	194	256
†Success	34	58	117	83	112	190	134	232	292	83	165	233	122	154	196
†Percentage Success	42	44	54	66	60	63	83	82	72	53	69	66	82	79	77

† See text. The "success" is that part of the variance of the observation that is successfully forecast, assuming that there are no errors in the observation.

TABLE II - Frequency of Errors in Forecast Temperatures.

Temp. °F.	AUGUST			SEPTEMBER			OCTOBER			NOVEMBER			DECEMBER		
	Frequency %			Frequency %			Frequency %			Frequency %			Frequency %		
	750	500	300	750	500	300	750	500	300	750	500	300	750	500	300
0-2	52	52	43	30	46	37	53	51	31	54	44	32	62	42	40
3-4	33	31	24	37	25	20	21	25	26	18	18	20	23	32	21
5-6	8	13	16	25	17	25	8	9	14	20	16	18	7	9	21
7-8	6	2	9	3	3	3	11	6	17	7	12	12	3	9	7
9-10	-	1	4	5	7	3	6	3	5	1	5	10	4	5	4
11-12	1	1	2	-	-	5	-	5	3	-	1	6	1	1	3
13-14	-	-	-	-	-	-	1	-	2	-	1	1	-	-	3
15-16	-	-	2	-	-	3	-	1	1	-	3	-	-	2	-
17-18	-	-	-	-	2	2	-	-	-	-	-	1	-	-	-
19-20	-	-	-	-	-	-	-	-	1	-	-	-	-	-	1
Probable Error	2·2	2·3	3·3	2·9	3·2	4·6†	3·0	3·3	4·2	2·6	3·6	3·7	2·4	3·2	3·7

† 2% greater than 20°F.

TABLE III - Frequency of Magnitude of Vector Errors in Forecast Winds.

Knots	AUGUST			SEPTEMBER			OCTOBER			NOVEMBER			DECEMBER		
	Frequency %			Frequency %			Frequency %			Frequency %			Frequency %		
	750	500	300	750	500	300	750	500	300	750	500	300	750	500	300
0-4	28	12	3	9	5	2	12	8	-	10	12	3	11	7	4
5-9	27	33	10	39	15	7	28	6	3	21	14	9	27	12	10
10-14	33	21	20	30	26	16	20	25	9	23	16	10	25	15	18
15-19	4	16	18	9	18	11	20	20	16	23	18	13	21	20	6
20-24	6	4	11	9	16	11	9	11	8	8	15	5	11	19	12
25-29	1	8	15	4	9	9	5	8	16	11	10	13	1	7	13
30-39	1	4	13	-	7	26	3	10	19	4	10	19	4	14	18
40-49	-	1	3	-	2	9	3	6	14	-	5	14	-	5	10
50-59	-	-	3	-	2	5	-	3	3	-	-	10	-	1	3
60-69	-	1	1	-	-	2	-	-	3	-	-	2	-	-	2
70-79	-	-	3	-	-	2	-	-	6	-	-	-	-	-	3
80 or over	-	-	-	-	-	-	-	3	3	-	-	2	-	-	1
Probable Error	7	11	19	9	14	21	11	19	28	11	14	23	8	15	22
Standard Error	11	16	28	13	21	31	17	28	41	16	21	34	15	22	32
12-hour Variance	13	18	35	19	24	40	23	34	51	23	29	40	18	27	38
†Success	7	9	20	14	13	25	15	20	30	16	19	22	10	15	20
†Percentage Success	55	46	59	74	52	63	68	59	59	69	68	54	56	55	54

† See footnote to Table I.

II. DISCUSSION OF THE RESULTS FOR AUGUST - DECEMBER, 1944.

As a basis for discussing the results for the last five months of 1944, the mean figures for percentage success over the period in question are given in Tables IV and V.

TABLE IV.

Mean Percentage Success for All Levels.

	Aug.	Sept.	Oct.	Nov.	Dec.	Whole Period.
Height	47	63	79	63	79	66
Wind Vector	53	63	62	64	55	59

TABLE V.

Percentage Success for Period August - December.

	750 mb.	500 mb.	300 mb.	All Levels.
Height	65	67	66	66
Wind Vector	64	56	58	59

The figures show that statistically the variations in height can be foreseen with rather more accuracy than the variations in wind. This might have been expected, since there appears to be no reason to suppose that the slopes of pressure surfaces can be predicted more accurately than the heights. Predicted winds are derived from the slopes of pressure surfaces together with consideration of the curvature effect. Furthermore owing to the complete absence of upper wind observations to the West of Larkhill, no allowance can be made for observed departures from the gradient wind.

There is no marked difference in the percentage success at the three levels. This is so in spite of the relative paucity of 300 mb. observations, particularly to the West of Larkhill. It may be interpreted as indicating that the properties of the 500 to 300 mb. layer are sufficiently known to ensure that, once the 500 mb. surface forecast has been considered in detail, no important fresh errors are introduced at 300 mb. if the layer by layer method of forecasting is adhered to.

The percentage success of the height forecast shows considerable variation from month to month, with maxima in October and December and a minimum in August. Thus the minimum success was attained in the month when the variance was least, while one of the maxima was scored in the month (October) when the height variances were greatest. It appears that in general the more pronounced the height variations become, the easier they are to forecast. This is bound to be the case if the forecast errors are of the same magnitude as the errors of observation. It is unlikely, however, that the errors in observation are comparable with the errors in the forecasts for August, so that the latter must be assumed real and the variations attributed largely to inhomogeneities in the atmosphere of relatively small extent which escape the network of soundings. In winter, on the other hand, the variations are mainly determined by fast-moving systems with pronounced patterns of upper contours, so that a larger fraction of the height fluctuations can be foreseen even though observations in the Atlantic are less closely spaced than over the British Isles.

Apart from the minimum success in August, the forecasts of wind from month to month do not show the same features. This is due to the fact that no wind measurements are available to the West of Larkhill. Thus the spacing of soundings in the Atlantic, while it suffices to reveal the pattern of upper contours, is not close enough to enable the strong gradients to be located with accuracy. As an example which occurs very frequently, the approach of a warm front from the Atlantic in winter is heralded by a sudden large increase and veer in the wind at upper levels. The forecaster expects this, but normally has no means to judge the exact magnitude of the maximum wind nor the precise distance of the maximum in advance of the front, which may vary from 200 to 700 miles. Frequently no reliable estimate of the maximum can be made until it is actually observed at Larkhill or Liverpool. Thus what is technically a good and sound forecast may yet show a large error if the forecast and actual winds at a given place and time be compared. An error of a few hours in timing in such situations may result in a vector error of well over 50 knots. Such errors will only be reduced when regular wind measurements from the Atlantic, or to a lesser extent from S.W.Ireland, are available. This type of error is eliminated from forecasts of upper wind for Western Europe, so that the accuracy of the Larkhill forecasts is by no means a perfect guide as to the precision of winds predicted for navigation to the East of the British Isles.

Little can be said of the errors made in forecast of temperatures except to draw attention to the smallness of their magnitude. The means of the probable errors for the five months were :-

750 mb.	500 mb.	300 mb.	
2·6	3·1	3·9	°F.

by authority of M.R.P. 305 12 Oct
by [illegible] date 12 Apr 59

TECHNICAL REFERENCE FILES

~~RESTRICTED~~ AIR MINISTRY MRP 227
26th March 1945

METEOROLOGICAL RESEARCH COMMITTEE.

Accuracy of Gregory Humidiometer,

by O.M. Ashford.

1. **Introduction.** The Gregory humidiometer is used in the meteorological equipment at Rye, designed to provide accurate records of the gradients of temperature and relative humidity between the ground and a height of 350 ft. In the present paper the Gregory humidiometer is discussed in detail with a view to assessing the accuracy of the records obtained from it and its general usefulness for meteorological work.

2. Gregory Element.

The original element has already been described in section 3 (a) of J.M.R.P. No.17. It has recently been reported that more stable elements can be made by using platinised steel for the rod electrodes instead of silver. This is presumably because of the better weather resistant properties of the platinised steel. It has been suggested that gold or a rhodium plated metal would be better still, but these have not been tried yet.

The simple theory of this type of hygrometer element is adequately covered in a recent paper by Schaffer (Schaffer 1945).

3. Laboratory Tests.

A full account of a series of tests carried out in the Meteorological Office Instruments Branch is given in Appendix 1, The chief results are summarised below:-

(a) Stability of Calibration. The element was almost unaffected by prolonged exposure to moist unsaturated air or by prolonged aspiration at 24 mph, but a definite permanent change of 2 - 3% was produced by exposure to wet fog for 24 hours.

(b) Lag. Earlier tests reported in J.M.R.P. No.17 indicated that the element responded to changes of humidity according to an exponential law, but more accurate tests have since shown that this is only a first approximation, as the element responds more rapidly than the exponential law, as it approaches the true humidity of the surrounding air. The tests do not cover a wide enough range of humidities to enable one to calculate the response law with confidence; the results obtained agree very closely with the empirical equations:-

$$\frac{dH}{dt} = \frac{-1}{0.6\lambda}(H - H')^{0.6} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (1)$$

$$\text{and } \lambda = \lambda_0 + \frac{193.5}{V^{1.15}} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (2)$$

where
H = indicated humidity
H' = actual humidity
V = speed of ventilatinn in m.p.h.
λ_0 is apparently a constant, equal to 3.9, for the range of humidity covered, but is probably a function of relative humidity and temperature.

Response curves obtained from actual measurements are shown in Figure 13, together with points calculated from equations 1 and 2.

/4.

4. Observations at Rye.

A report on the performance of the Gregory humidiometer at Rye is given at Appendix 2. The main conclusions are:-

(a) Stability of Calibration. Apart from sudden changes due to exposure to wet fog, the elements are fairly stable, showing a gradual change in calibration of only 2 to 3% after 6 months operation.

(b) Lag. The naturally small lag of the elements is often artifically increased owing to inadequate ventilation of the screens. Errors up to 14% persisting for an hour or more may occur. These results, which were obtained by comparison with observations with Assmann psychrometers, have subsequently been verified by records obtained from an element mounted in a screen artifically aspirated by an electric fan at a speed of about 8 m.p.h. This will be the subject of a separate report.

5. Review of Results.

(a) Accuracy.

The Gregory elements are designed to be used with a standard chart or dial, without the application of individual corrections. Corrections have to be applied to compensate for the effect of temperature on the resistance, but this is done by a standard graph which is intended to apply to all elements. (see JMRP No.17). The apparent errors in new elements found by comparison with accurate psychrometers are generally less than ± 4%. The curves of apparent error against relative humidity for various elements show considerable parallelism, the mean error varying from + 1% at 30% to - 2% at 70% relative humidity. This systematic difference may be accounted for by differences in the calibration techniques used by the makers and the users, i.e. between solution hygrometry and wet bulb hygrometry.

Random errors of the psychrometers under laboratory conditions are estimated at ±1%, but at Rye additional allowance must be made for the true difference between the air sampled by the psychrometer and that surrounding the element; the random errors are largely overcome by meaning several observations.

It is concluded that a new Gregory element will give readings of relative humidity to within ± 3% without the application of corrections other than for temperature, and to ± 1% if corrections are obtained from regular comparisons with an Assmann psychrometer. The uncorrected readings might be 6% in error after 6 months exposure under normal conditions, and much greater if the element had been subjected to wet fog during the period.

(b) Lag. Although equations (1) and (2) are only based on a limited number of observations over a limited range of humidity and temperature, it seems fair to state that the response of the Gregory element has two components, one dependent on the speed of ventilation and the other independent. These can probably be identified as the lag due to diffusion resistance at the boundary layer and that due to internal diffusion respectively, in the same way as Glückauf has observed with gold-beater's skin (Glückauf 1943). The latter component, λ_o, sets a limit to the rate of response for increased speeds of ventilation, and very little improvement in response would be obtained by increasing the ventilation above 100 m.p.h. At room temperatures and for humidities between 45% and 85%, λ_o is 3.9, so that under these conditions the time required for the element to indicate 50% and 90% of a sudden change of 40% in the relative humidity would be 13 seconds and 28 seconds for a ventilation speed of 100 m.p.h. For an infinite ventilation speed, the latter figure would only be reduced to 26 seconds. For most applications on the ground, a ventilation speed of 10 m.p.h. would be ample, but for measurements on aircraft it would be desirable to have a ventilation of about 100 m.p.h. As for all other hygrometers based on the absorption principle, the lag of the Gregory element is bound to increase very rapidly with decreasing temperature, and as its lag under ground conditions is only slightly better than that of human hair and much worse than gold beaters' skin, it offers very little hope of development for use as a satisfactory aircraft or radio-sonde hygrometer at low temperatures. The lags of various other hygrometers at room temperature for a sudden change of humidity from 85% to 45% have been estimated from various published results and tests made in the Meteorological Office, and are given in the following table:-

Speed of Ventilation.	6 m.p.h.		100 m.p.h.	
Hygrometer	50% of change secs.	90% of change secs.	50% of change secs.	90% of change secs.
Gregory.	88	185	13	28
Gold-beaters' skin.	4	12	0.9	4.3
Hair (Nyberg).	26	86	26	86
" (M.O. hygrometer)	70	200	-	-
" (N.& Z. baro-thermo-hygrograph)	30	140	-	-
" (Experimental Aircraft Hygrometer).	20	250	30	100
" (Dunmore)	30	240	-	-
Bureau of Standards Electrical (Dunmore).	2	18	-	-

The difference between the Gregory element and the Bureau of Standards electrical hygrometer, which is very similar in principle (Dunmore 1939), is rather surprising. Details of the experimental determinations of lag in the Bureau of Standards are not given in the report from which the above lag figures were obtained (Dunmore 1940), but differences of technique could only account for a small percentage of the difference in lag coefficients. It can be concluded that the Gregory element is by no means the most rapidly responsive of the hygrometers using lithium chloride as the hygroscopic medium.

(c) Range.

Owing to the marked non-linearity of the calibration, it is only possible to cover a limited range of relative humidity adequately with a single element. At Rye for example it has been found necessary to install a second set of elements to cover the low humidities (below 35%) occasionally experienced there. It is possible that the disadvantages of non-linearity could be partially overcome by having two or more elements, each with maximum sensitivity at different parts of the humidity scale, connected up in parallel, as has been suggested by Dunmore (See BCSO Report No. 290) for National Bureau of Standard Electrolytic hygrometers.

6. Conclusion.

The chief merits of the Gregory element are its great sensitivity over limited ranges of relative humidity and its ready application to distant indication and recording. For such a sensitive element working on the absorption principle its stability, while leaving room for improvement, is reasonably satisfactory.

The main disadvantages are the non-linearity of the calibration, the large variations with temperature, the need for protection against direct contact with water droplets and the slow rate of response at low temperature.

No tests have been made to determine the behaviour of the element under tropical conditions, but it is suggested that a fabric made from an inorganic material, such as "Eutex" glass fibre, might give better results in the tropics, as it should be more resistant to the growth of fungus.

/Summing

Summing up, it seems that the Gregory element can be used as a satisfactory hygrometer for the following applications:-

(a) where changes of less than 1% have to be detected over limited ranges.
(b) where an accuracy of 1% is required, subject to checks against an Assmann psychrometer or other standard hygrometer.
(c) where automatic recording is required.
(d) where distant indication or recording is required.

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W. Schaffer 1945. Simple theory of the electric hygrometer, S.A.A.F. Met. Section. Technical Notes No.37.

APPENDIX I.

Report on the Gregory Humidiometer Laboratory Tests.

by

N.S. SMITH.

I. Apparatus.

1. The portable humidiometer.

The arrangement of the circuit is shown in Figure 12and the indicator for this apparatus in Figure 2. A single transformer reduces the mains voltage from 230 to 11 and the secondary is connected across a circular 500 ohm potentiometer. Three positions of the main 5 pole 3 way switch give the following:-

(a) Low Range relative humidity. The 500 ohm potentiometer is tapped directly through a resistance of 10,800 ohms, the A.C., 0-500 microammeter and the element connected in series, the microammeter being calibrated from 25% R.H. to 50% R.H.

(b) High Range relative humidity. In this case, the 500 ohm potentiometer is tapped through the primary of a step up transformer (about 1-10), a 550 ohm resistance and the element connected in series. The secondary of this transformer across which there is a potential of about .4 volts is connected directly through the A.C. microammeter.

(c) Voltage check. In this position of the switch the connections are the same as for the high range R.H. except that a 52 ohm resistance takes the place of the element. The tapping of the 500 ohm potentiometer is adjusted to give a full scale reading on the microammeter.

A 12 v. accumulator can be used instead of the 230 v. mains by means of a vibrator converter incorporated in the humidiometer, and the 500 ohm potentiometer is fed either by the output of the converter or by the secondary of the step down transformer according to the position of a D.P.D.T. **switch.**

At low humidities the actual voltage across the element rises to a maximum of about nine volts.

2. Humidity Test chamber (Kew Pattern).

This consists of a main chamber around which the air is circulated by means of a fan, and an air circulation system which allows for air at different humidities to be present in the main chamber.

Air travels from the compression side of an Edwards rotary pump to the bottom of the main chamber via a buffer cylinder and either (a) a cylinder containing silica gel, (b) a cylinder containing cold water, or (c) a cylinder containing warm water, depending on the position of a three way cock. On leaving the main chamber the air flows past the bulbs of a pair of Assmann thermometers, one of which has a wet bulb, and thence to the suction side of the pump.

Terminals are fitted to one of the lids of the chamber and these are used to connect the Gregory element electrically to the portable humidiometer.

II. Description and results of tests.

1. Calibration.

Elements No. 25 and No. 26 were received from Messrs. Negretti and Zambra,

/and

and Figs. 3 and 7 show a pair of calibration curves for each of the elements. The wick of the wet bulb was thoroughly cleaned with soap and water and finally rinsed with distilled water before each pair of these and later calibrations.

From these curves it may be reasonably deduced that:-

(a) The accuracy obtainable from the test apparatus is about ± 1% R.H.

(b) Element 25 requires a correction varying from -2% to +6% R.H. over the range 30% - 100% R.H.

(c) Element 26 requires a correction varying from -1½% to +2% R.H.

2. Stability of calibration.

(a) After prolonged subjection to unsaturated air at high humidity.

Element 25 was kept in the humidity chamber for 5 hours with the air maintained at 90 to 95% R.H., and figure 4 shows the calibration curves before and after this treatment.

Bearing in mind that the accuracy of the observation technique is about ± 1% R.H., it is observed that there is no appreciable change in calibration except perhaps a reduction of about 3% R.H. in the positive correction at 90% R.H., but this deduction is only supported by a single point on the graph.

(b) After prolonged subjection to wet air.

Element 25 was kept for 24 hours within a bell jar over water and with the roof of the bell jar covered with moistened cellulose wadding. In the evening the room became much cooler and fog was formed within the bell jar resulting in the element becoming quite wet. Figure 5 shows the calibration curves before and after this treatment.

It is observed that an increase of 2-3% R.H. in the positive correction has been caused. This doubtless due to the washing away of some of the lithium chloride.

(c) After subjection to prolonged aspiration at 24 m.p.h.

The element was kept for 4 hours in the air stream of a small wind tunnel, the wind speed being checked by an airmeter. Figure 6 gives the calibration curves before and after aspiration.

No substantial change in the calibration can be observed.

In all these stability of calibration tests the second calibration was made immediately after treatment and in cases (a) and (c) the first calibration was made in the afternoon preceding the morning in which the treatment took place. With case (b) the first calibration was made during the morning immediately preceding the 24 hours subjection to wet air.

It will be observed that the positive correction to element 25 has increased from about 1% R.H. to 7% R.H. during a period of three months.

Figure 3A shows the calibration of element 26 after a period of about 5 months, and here the positive correction has increased from about 1% R.H. to 6% R.H. This element was used for the lag tests.

The humidiometer was found to be rather insensitive from 80% R.H. to 100% R.H. and accurate readings could not be obtained in this region.

/III.

FIG. 1

GREGORY PORTABLE HUMIDIOMETER : INDICATOR.

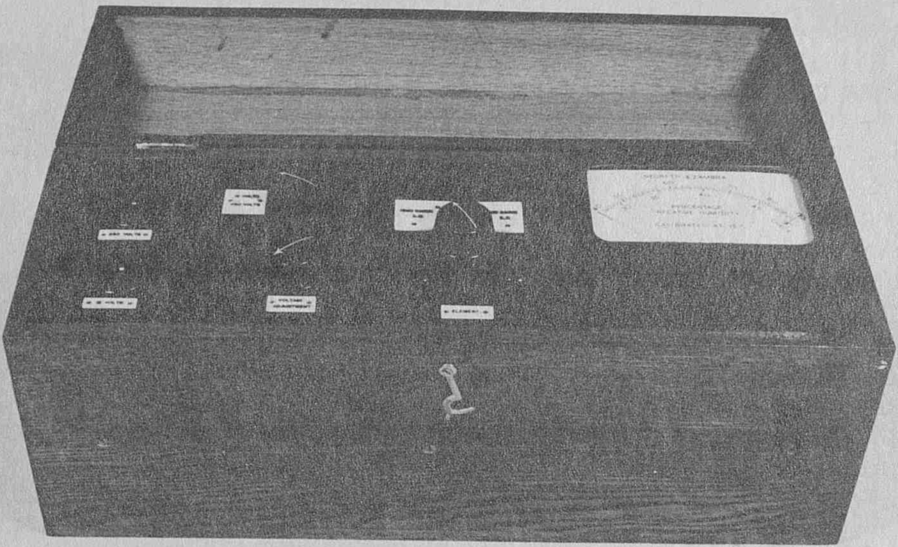
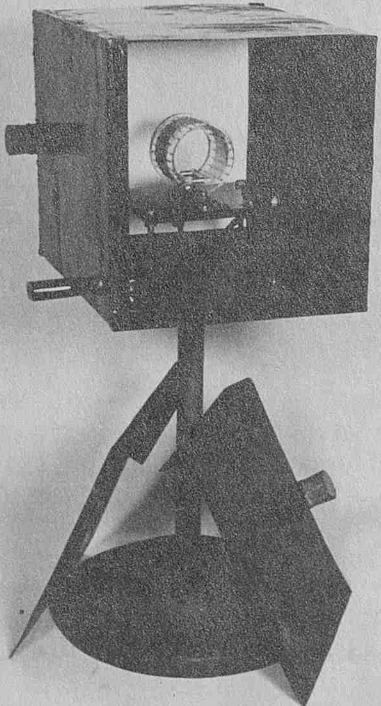


FIG. 2.

GREGORY ELEMENT IN HOUSING FOR LAG TESTS.



III. Lag Tests.

These were carried out on element 26 using the container shown at figure 2. The container was designed so that it could be placed in the humidity chamber with the sides in position, the corks extracted and a dummy Igranic P.40 plug at A. In this way a good proportion of the air in the chamber circulates round the element as shown in figure 8.

When the humidity chamber had been kept at about 95% R.H. for about ten minutes, care being taken to avoid saturation at 100% R.H. since the moisture drops would produce an abnormal lag, the air circulation was stopped, the lid of the chamber removed and the element container corked up quickly at C. and D. The container was then taken to the wind tunnel room where the motor was running at the speed necessary to produce the required airflow, and the portable humidiometer was being operated by a 12 volt accumulator, the element terminals being connected externally to two terminals fitted to the floor of the wind tunnel. These two latter terminals were connected within the tunnel to a second Igranic P.40 plug by a short length of flex. The element container was placed quickly on its stand at the centre of the wind tunnel, the dummy plug was replaced by the second plug, and after removing the sides of the container the door of the wind tunnel was shut. A stop watch was started when the humidiometer reading had reached a multiple of 5, and subsequent readings of the stop watch were taken for each fall of 5% R.H. The observations were set out in the following example:-

$H_\alpha = 47.0°F.$

t (sec)	H	$H-H_\alpha$	$\log_e(H-H_\alpha)$	$\log_e(H_0 - H_\alpha) - \log_e(H - H_\alpha)$
0	85	38	3.6376	0
21	80	33	3.4965	0.1411
40	75	28	3.3322	0.3054
60	70	23	3.1355	0.5021
85	65	18	2.8904	0.7472
110	60	13	2.5650	1.0726
137	55	8	2.0794	1.5582
186	50	3	1.0986	2.5390

H = Humidiometer reading at time t.
H_α= Room relative humidity as given by the final steady reading.
H_0= Humidiometer reading after 0 sec.

$$\lambda = \text{lag coefficient} = \frac{t}{\log_e(H_0-H_\alpha) - \log_e(H-H_\alpha)}.$$

t was plotted against $\log_e(H_0 - H_\alpha) - \log_e(H-H_\alpha)$ and each lag test was carried out three times except at 40 m.p.h. when two tests were made. It will be noticed that the room relative humidity should be less than 50% for a reasonably good lag test to be made.

Figure 9 shows the first three lag tests at 6 m.p.h. using supports ½" long for fixing the element in the container. The lag is seen to be exponential and of average value 88 seconds. No temperature corrections were made in these and later lag tests since the effect would be negligible.

In order to obtain better aspiration, longer supports were now fixed to bring the element to the centre of the container and lag tests were carried out at wind speeds of 6 m.p.h., 25 m.p.h. and 40 m.p.h. According ~~to these~~ tests the lag is not exponential, and the time taken to fall 50% and 90% of the difference between the indicated humidity and the actual humidity is shown at Table I.

Table I.

Wind Speed.	Time to fall 50% (H - Hα)	Time to fall 90% (H - Hα)
6 mph (with housing)	229 sec.	520 sec.
6 mph	88 sec.	185 sec.
25 mph	29 sec.	55 sec.
40 mph	20 sec.	43 sec.

Lag Tests using a housing for the element

Three additional lag tests were carried out at a wind speed of 6 m.p.h. using a special element housing provided by Messrs. Negretti and Zambra and shown at figure 10. The direction of the wind was at right angles to and toward the pair of copper gauze windows. About 30 minutes had to be allowed for the element to assume the high humidity of the chamber prior to the lag tests, and stop watch readings were taken for falls of 2% and 3% R.H. alternately. The average value of the lag for these tests is included in Table I.

Figure 11 shows graphically the average results of the lag tests. Only a portion of graph D is included.

APPENDIX 2.

Report on Performance of Gregory Humidiometer at Rye.

by K. Stormonth and O.M. Ashford.

1. Stability of Calibration.

Frequent measurements of the relative humidity have been made with Assmann psychrometers for comparison with the values obtained from the Gregory humidiometer records. From a plot of the errors thus found, a mean curve of error against relative humidity was obtained, and a comparison of these curves over intervals of several months gives a measure of the stability of the Gregory element. The calibrations were done at times when the air around the elements was considered to be representative of the air outside the screen.

Results for two typical elements are shown in Figures 14 and 15. Both show that the reading of the Gregory humidiometer becomes progressively lower with time, the change over a period of six months being of the order of 2-3%.

Generally it has been possible to use an element for six months or more, but two cases have occurred of sudden deterioration. One of the original set of elements was badly damaged by wet fog in November, 1943. Water was deposited copiously on the element and streamed down the muslin carrying a considerable quantity of the hygroscopic material with it. A set of elements installed in November and December, 1943, showed the usual slow changes until the end of July, 1944, when a sudden decay occurred in all five elements. The decay was attributed partly to the effect of blast from a flying bomb explosion nearby.

2. Exposure of Elements.

Whereas under conditions of steady relative humidity and of sufficient wind to ventilate the screens comparisons with Assmann readings showed that, subject to a small correction based on the curves described in Section 1 above, the results obtained from the Gregory humidiometer records could be regarded as an accurate indication of the conditions in the free atmosphere, there have been numerous occasions on which the records failed to give such a true picture.

Analysis of a number of such occasions shows that the most serious cases can be roughly divided into five classes:

(a) Commencement of precipitation. There is a lag in the rise of relative humidity around the element compared with the air outside. The lag varies according to the intensity of the precipitation and the wind force. Thus during thundery showers with little wind, discrepancies of 10 to 14% may arise. The effect may continue for as long as 20 to 30 minutes. See examples 1 and 5.

(b) Nocturnal cooling. When the relative humidity rises as a result of nocturnal cooling, the air temperature around the element will fall less rapidly than the outside temperature when the wind is light. Under such conditions the recorded relative humidity may be 5 to 7% too low. See example 2.

(c) Cessation of precipitation. After precipitation, the presence of water on and in the Stevenson screen causes the element to read too high. This is especially so if the external relative humidity falls rapidly as a result of solar heating after the disappearance of the clouds. The error may be as large as 14% and may persist for up to 2 hours. See examples, 1, 3, 4 and 5.

(d) Clearance of Fog. If fog disperses as a result of solar radiation, the temperature inside the housing lags behind that outside, and the humidity reading will be too high, especially if there is little wind. Sometimes after a calm night with nocturnal cooling giving rise to effect (b), the wind may increase considerably with sunrise, and the element may read higher for a time than at any stage during the night. See example 2.

(e) Wet fog in still air. As mentioned in section 1, the heavy deposition of moisture during wet fog may cause permanent changes in the characteristics of the Gregory element.

The effects described in paragraphs (a) to (d) are essentially of a temporary nature, and may all be ascribed to insufficient ventilation of the humidity element.

Example 1. 10th February, 1944. Following the commencement of snow showers, the 4 ft. record showed an increase from 70% to 83% between 1400 and 1412, compared with an Assmann reading of 88% at 1410. After the snow, an Assmann reading of 69% was obtained at 1500, but the recorded humidity did not fall to 69% until 1530. The recorded humidity was thus some 6% low after the beginning of the precipitation and high by about the same amount after the cessation.

Example 2. 10th March, 1944. Fog formed at 0500 after a slight mist overnight, and did not clear until 0900. The recorded humidity was only 86% at 0800, when it was found that the air inside the screen was 1°F. warmer than the air outside. Calm conditions prevailed. An Assmann reading at 0900 gave 94%. After the clearance of the fog the element read 3% too high at 1000 and 4% too high at 1100. It was reading correctly by 1200. The wind had risen from 2 m.p.h. at 0900 to 10 m.p.h. at 1100.

Example 3. 2nd July, 1944. There had been rain and drizzle the previous night, and during the morning the 350 ft. element, which was in low stratus cloud and ventilated by a 10 m.p.h. breeze, recorded correctly about 99%. By 1100 the drizzle ceased and Assmann readings at 1330 and 1430 were 90% and 79%[x] respectively. This illustrates the ability of the elements to record high humidities correctly when adequately ventilated, and also the subsequent lag due to the wet screen even in a 10 m.p.h. breeze.

Example 4. 28th August, 1944. Rain commenced at 0830 and ceased at 0912. Recorded values and Assmann readings at 0845, 0855, 0930, 1000, 1030 and 1100, showed that the elements were reading 1%, 5%,6%,6%,4%, and 10% too high respectively. Good agreement was reached by 1200. There was a moderate breeze throughout. This again illustrates the prolonged effect of water upon the louvres of the screen.

Example 5. 3rd September, 1944. At 1300 the recorded humidity was 60.5% compared with an Assmann reading of 61%. Following a few showers at 1335, the record showed a steady increase to 71.5% at 1345, whereas an Assmann reading at 1340 gave 83%. At 1400 the recorded value of 69% was however, 3% too high, but it was correct by 1420. This shows the lag after the commencement of precipitation and a lag in the opposite direction after the cessation.

[x] whereas the recorded values were 100% and 87%.

MRP 227

FIG. 3

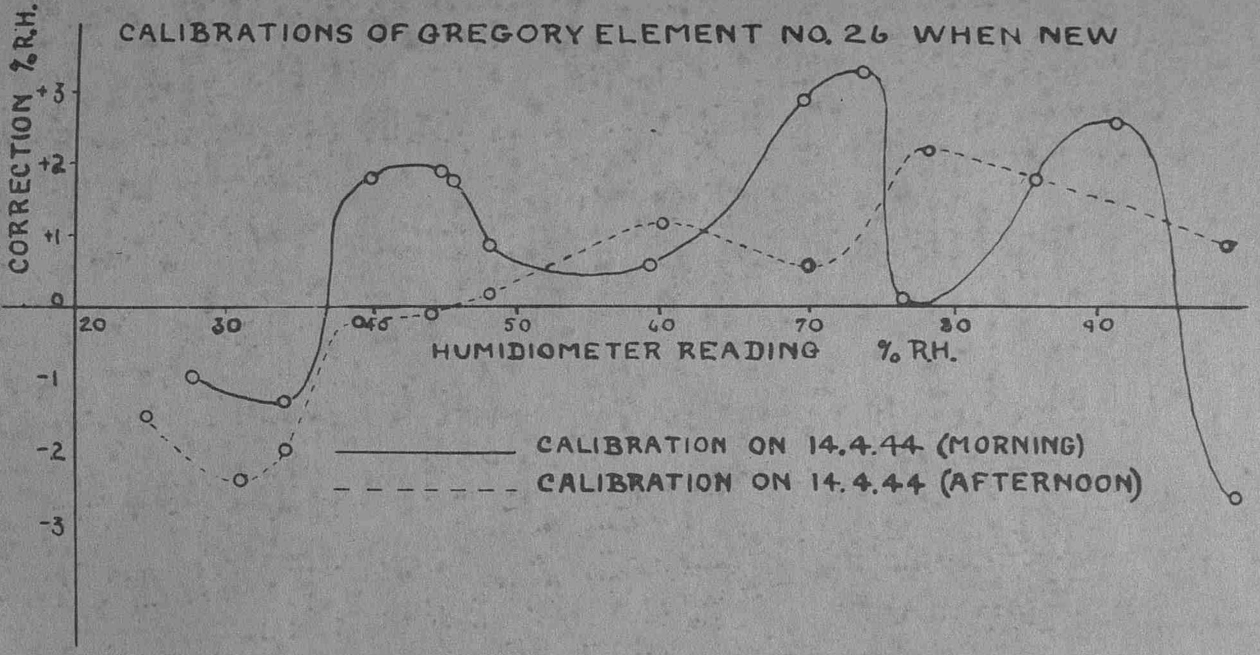
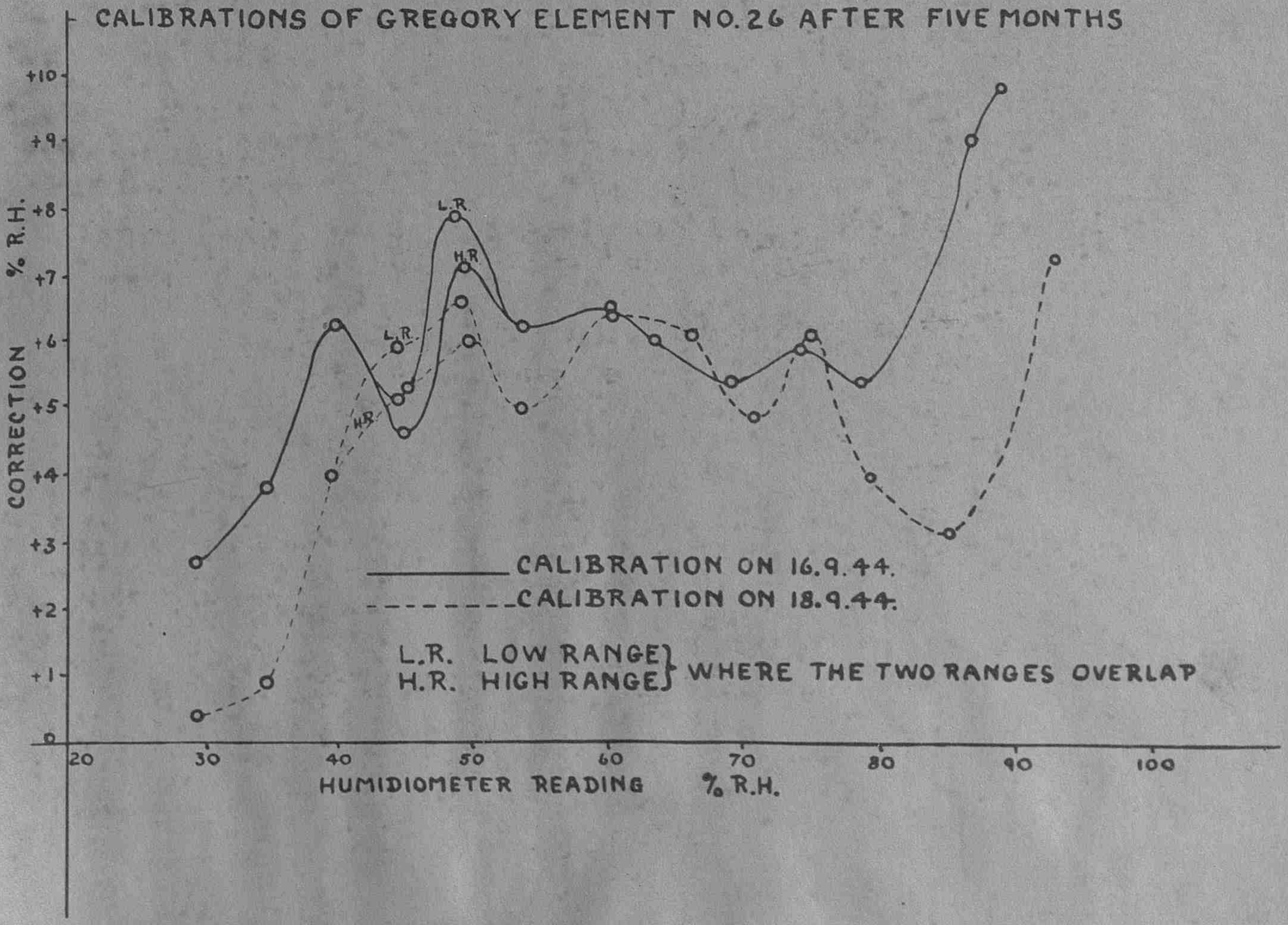
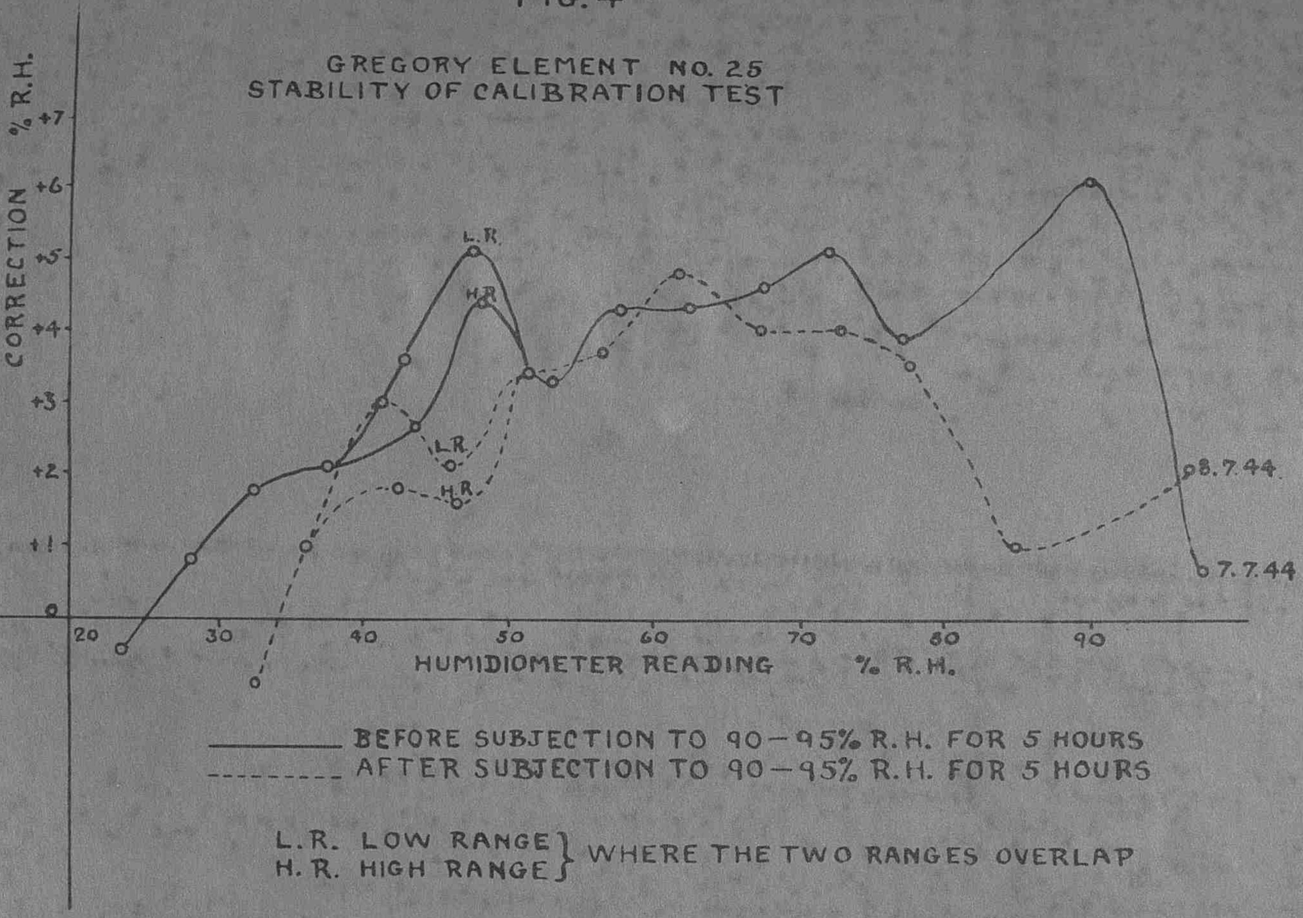

CALIBRATIONS OF GREGORY ELEMENT NO. 26 WHEN NEW

FIG. 3A


CALIBRATIONS OF GREGORY ELEMENT NO. 26 AFTER FIVE MONTHS

MRP227

FIG. 4



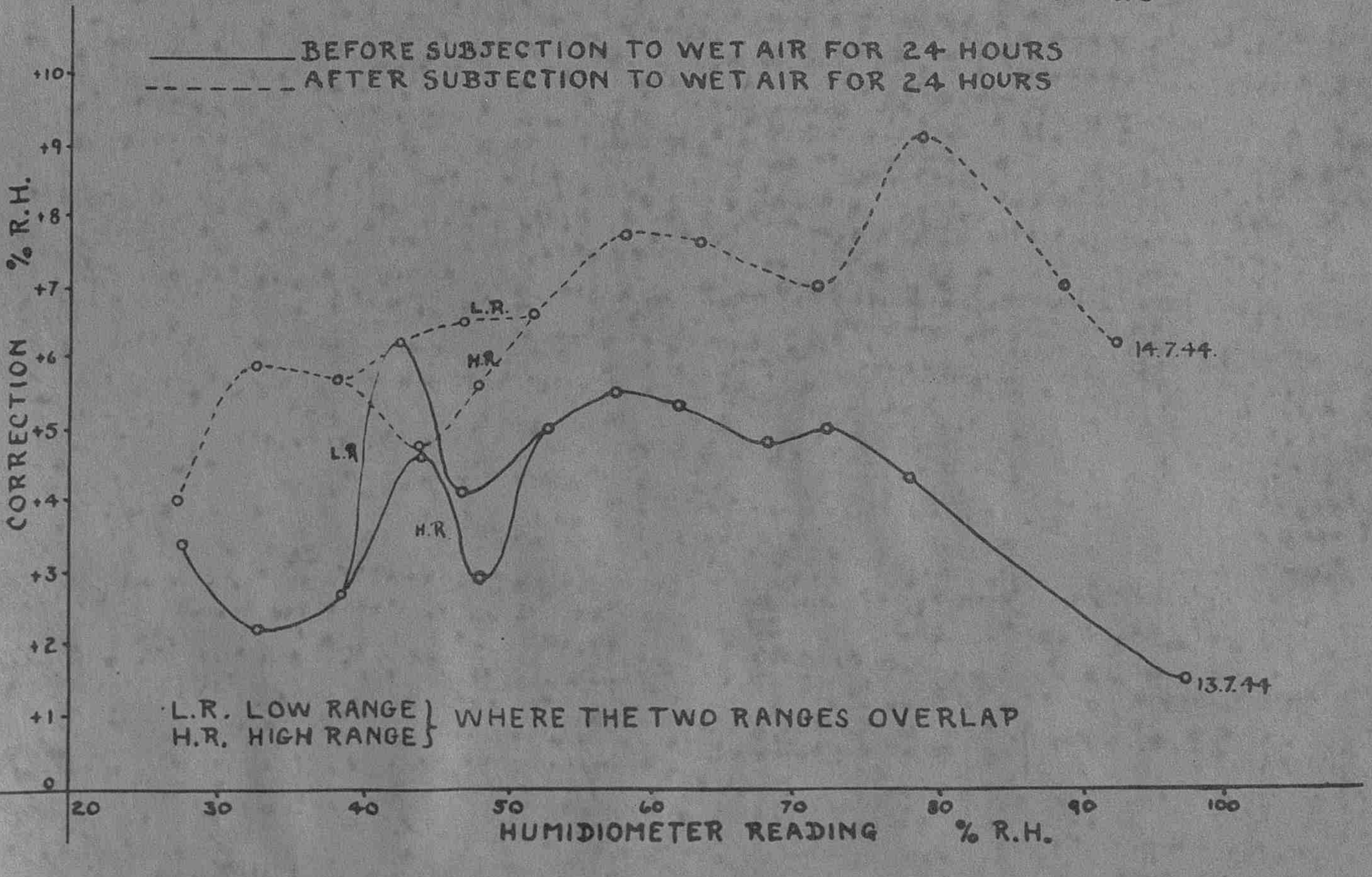
GREGORY ELEMENT NO. 25
STABILITY OF CALIBRATION TEST

———— BEFORE SUBJECTION TO 90–95% R.H. FOR 5 HOURS
- - - - - AFTER SUBJECTION TO 90–95% R.H. FOR 5 HOURS

L.R. LOW RANGE }
H.R. HIGH RANGE } WHERE THE TWO RANGES OVERLAP

FIG. 5

GREGORY ELEMENT NO. 25. STABILITY OF CALIBRATION TEST



———— BEFORE SUBJECTION TO WET AIR FOR 24 HOURS
- - - - - AFTER SUBJECTION TO WET AIR FOR 24 HOURS

L.R. LOW RANGE }
H.R. HIGH RANGE } WHERE THE TWO RANGES OVERLAP

FIG. 6

GREGORY ELEMENT No 25.
STABILITY OF CALIBRATION TEST.

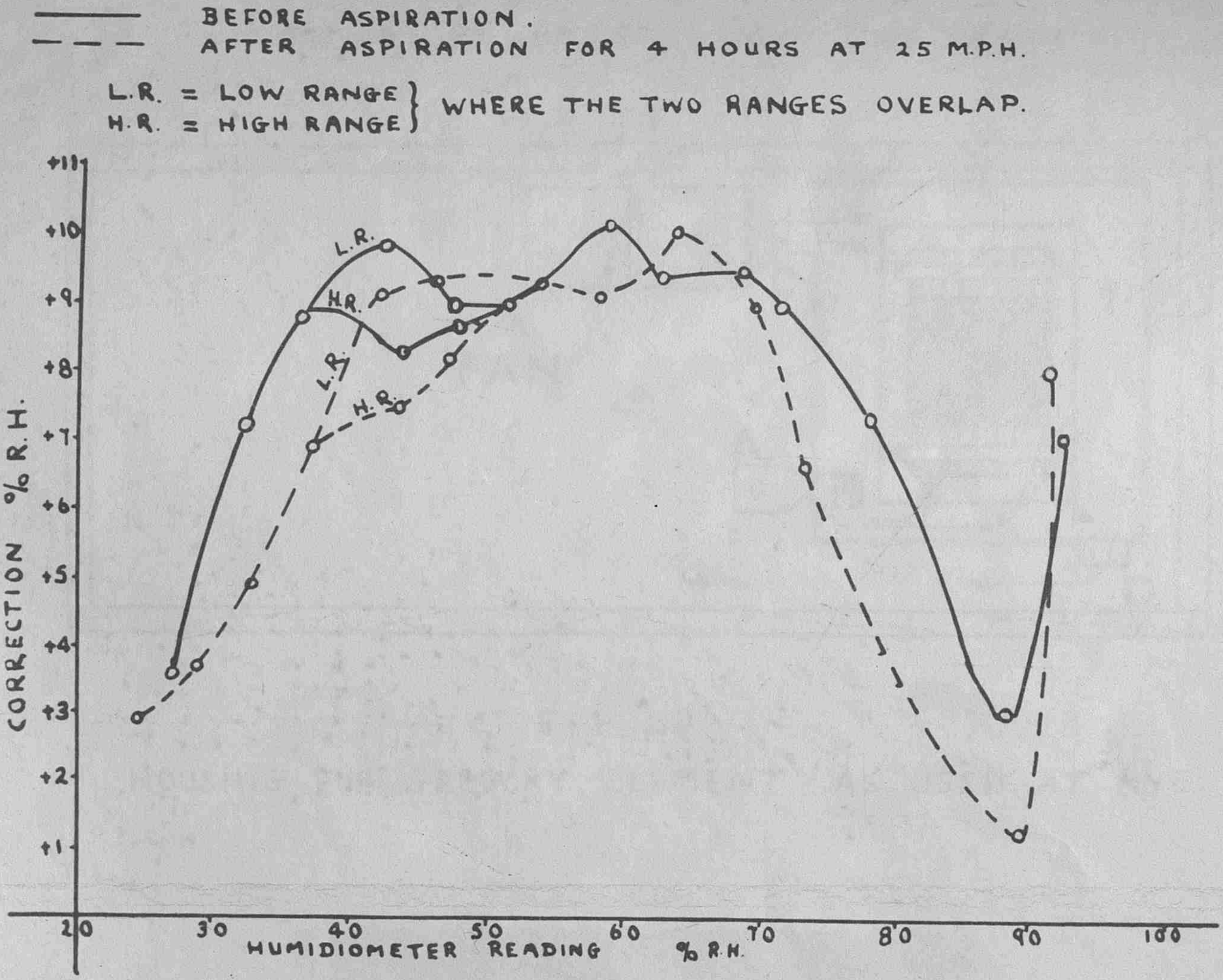


FIG. 7

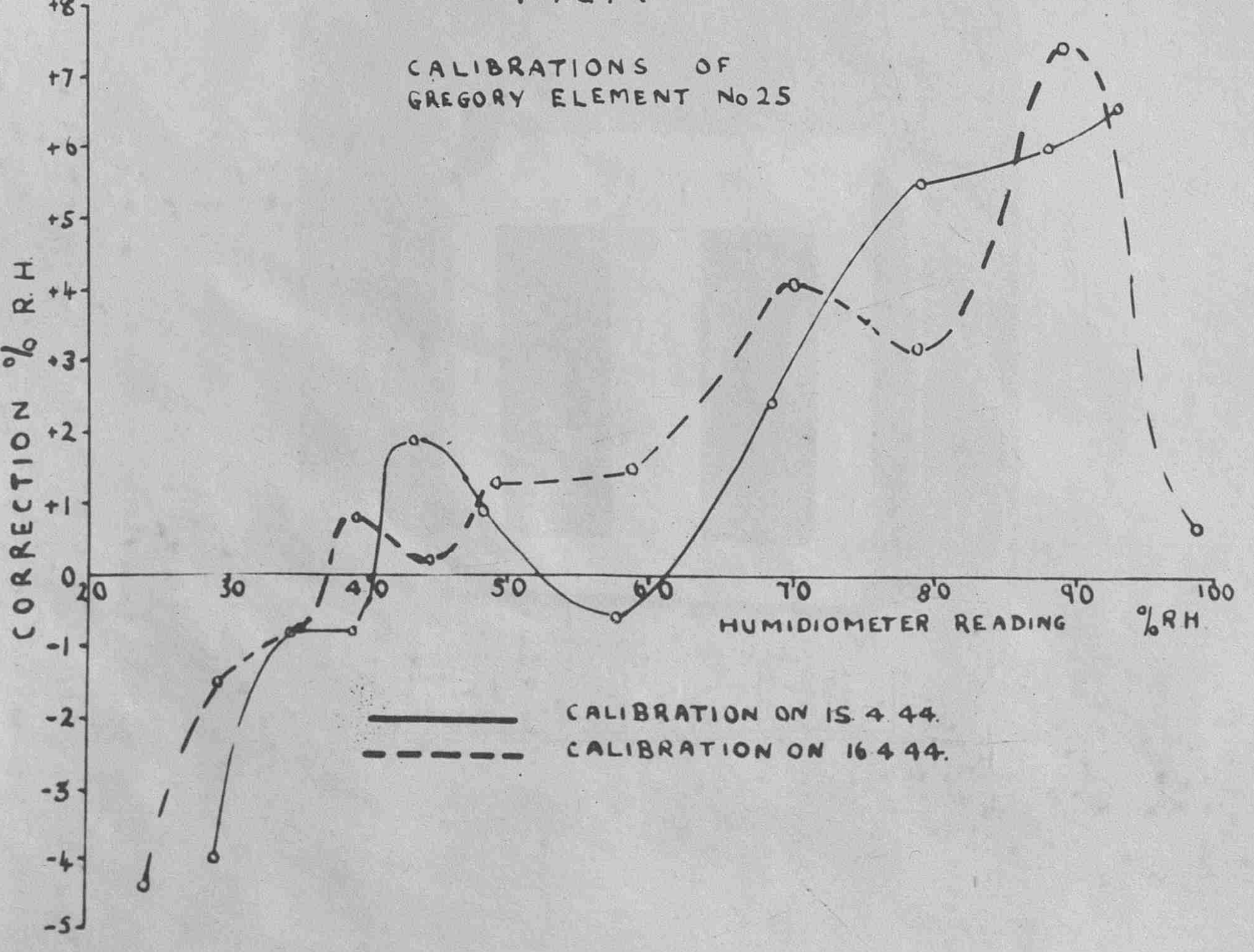

CALIBRATIONS OF GREGORY ELEMENT No 25

FIG. 8

HUMIDITY TEST CHAMBER WITH ELEMENT BEING PREPARED FOR LAG TESTS.

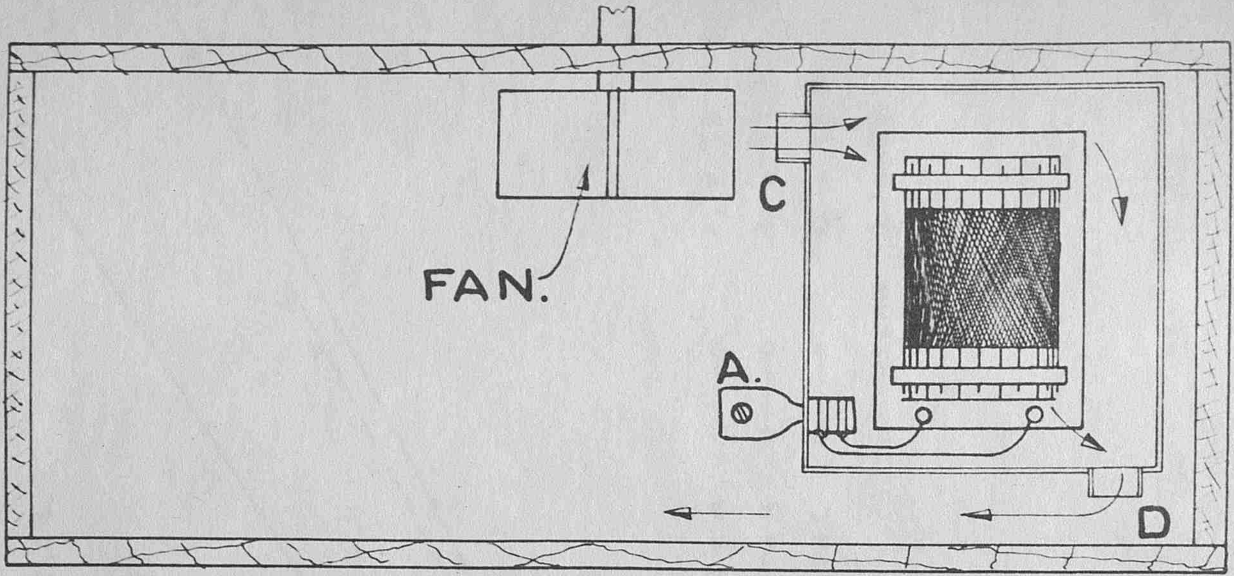


FIG. 10.

HOUSING FOR GREGORY ELEMENT AS USED AT RYE.

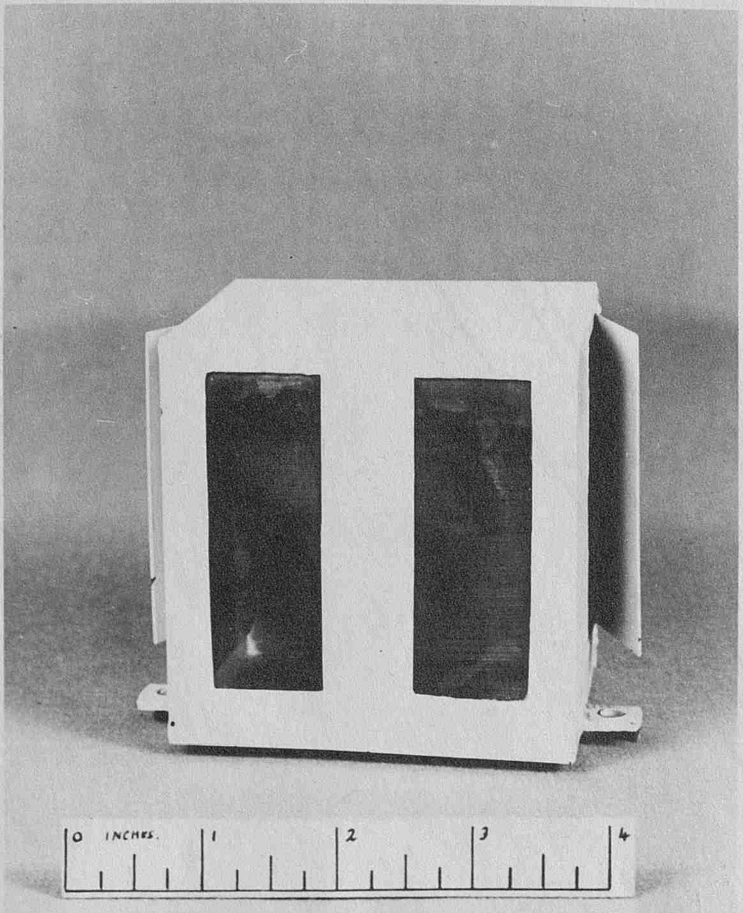

0 INCHES.
1
2
3
4

FIG. 9

GREGORY ELEMENT No. 26

LAG TESTS USING SHORT SUPPORTS 2.5.44.

WIND SPEED 6 M.P.H.

TEMPERATURE 62.0 – 62.3°F.

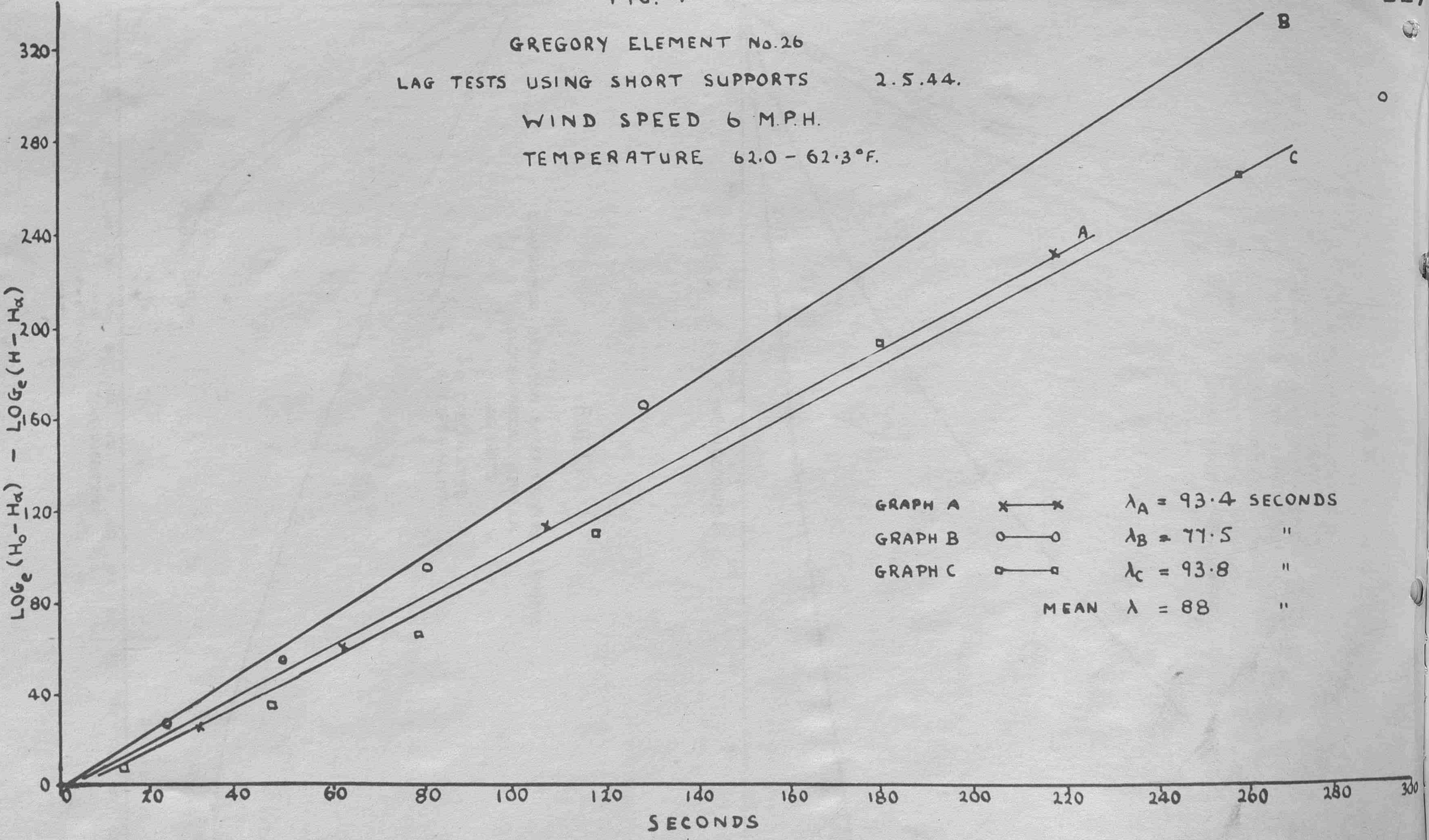
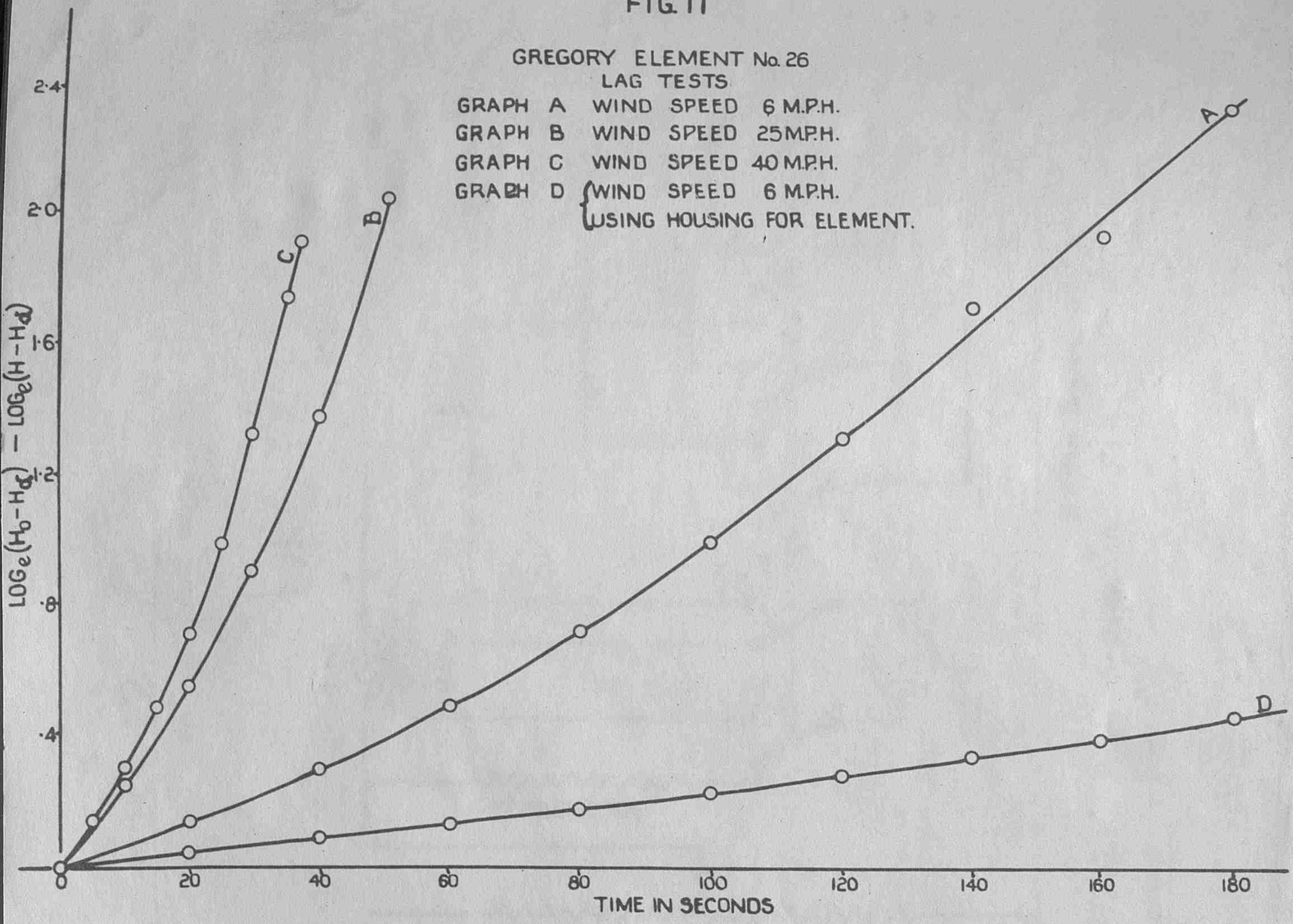
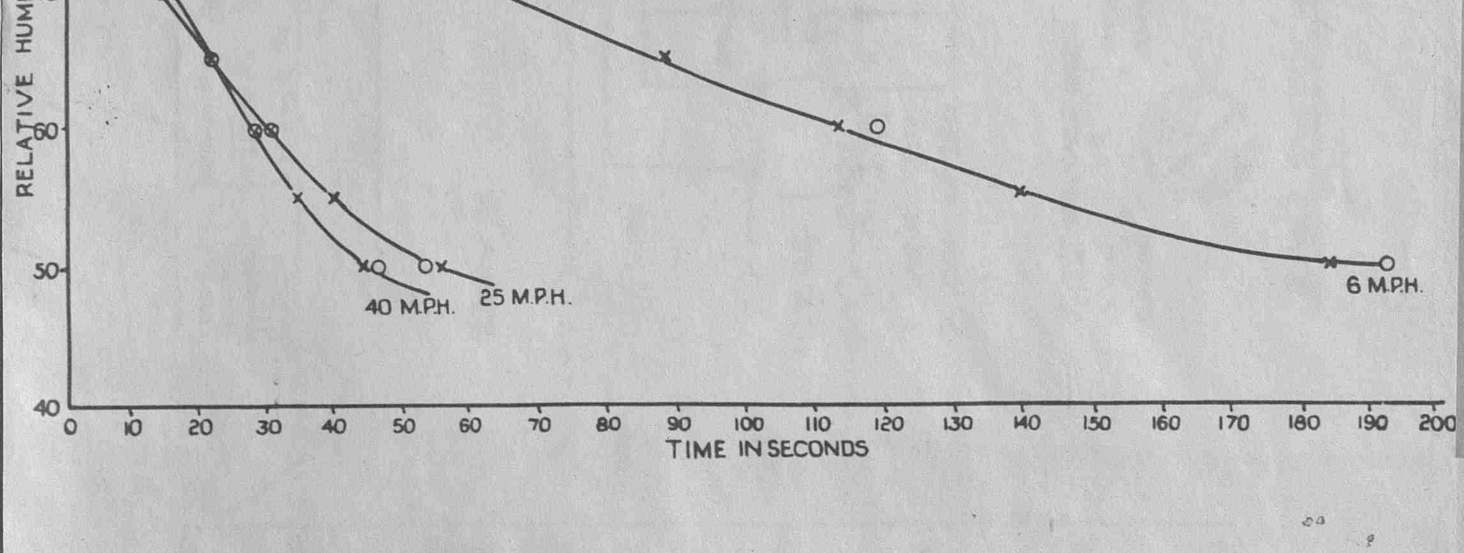


FIG. 11



GREGORY ELEMENT No. 26
LAG TESTS
GRAPH A WIND SPEED 6 M.P.H.
GRAPH B WIND SPEED 25 M.P.H.
GRAPH C WIND SPEED 40 M.P.H.
GRAPH D {WIND SPEED 6 M.P.H.
USING HOUSING FOR ELEMENT.

FIG. 13



COMPARISON BETWEEN EXPERIMENTAL RESULTS
AND EMPIRICAL FORMULA
LAG TESTS
O CALCULATED
X EXPERIMENTAL

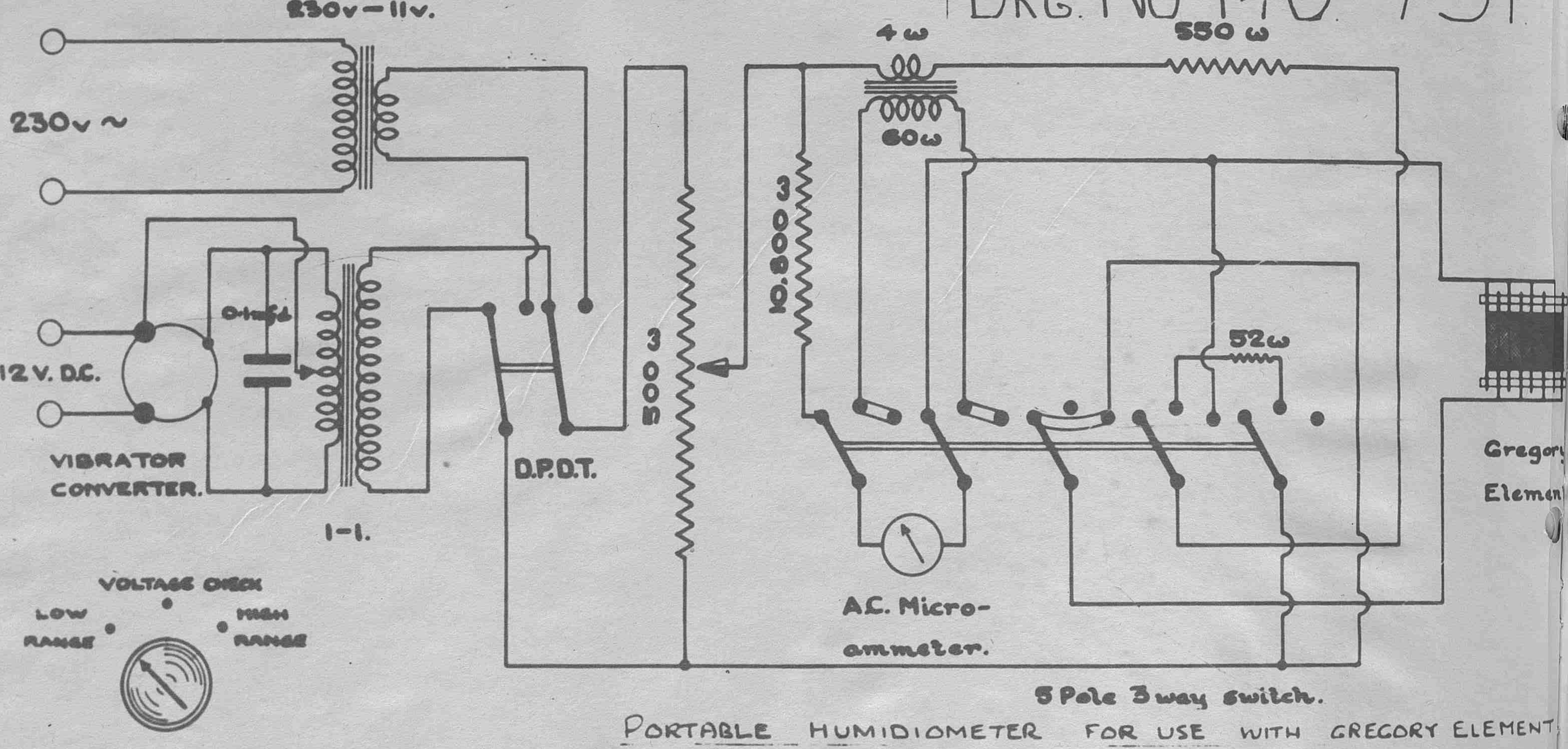
FIG. 12.

Issue 1
New drg
Approved.

Drawn By 22-9-44
Checked 22-9-44
AM File No E 945-43

AIR MINISTRY METEOROLOGICAL OFFICE

DRG. No MO 791


230v—11v.
230v ~
12 V. D.C.
0·1mfd
VIBRATOR CONVERTER.
1-1.
D.P.D.T.
500 ω
10,800 ω
4 ω
60 ω
550 ω
52 ω
A.C. Micro-ammeter.
Gregory Element
VOLTAGE CHECK
LOW RANGE
HIGH RANGE

5 POLE SWITCH KNOB.

5 Pole 3 way switch.

PORTABLE HUMIDIOMETER FOR USE WITH GREGORY ELEMENT

WIRING DIAGRAM.

FIG. 14

CALIBRATIONS OF 4 ft. GREGORY ELEMENT AT RYE
10. 12. 43 TO 1. 8. 44

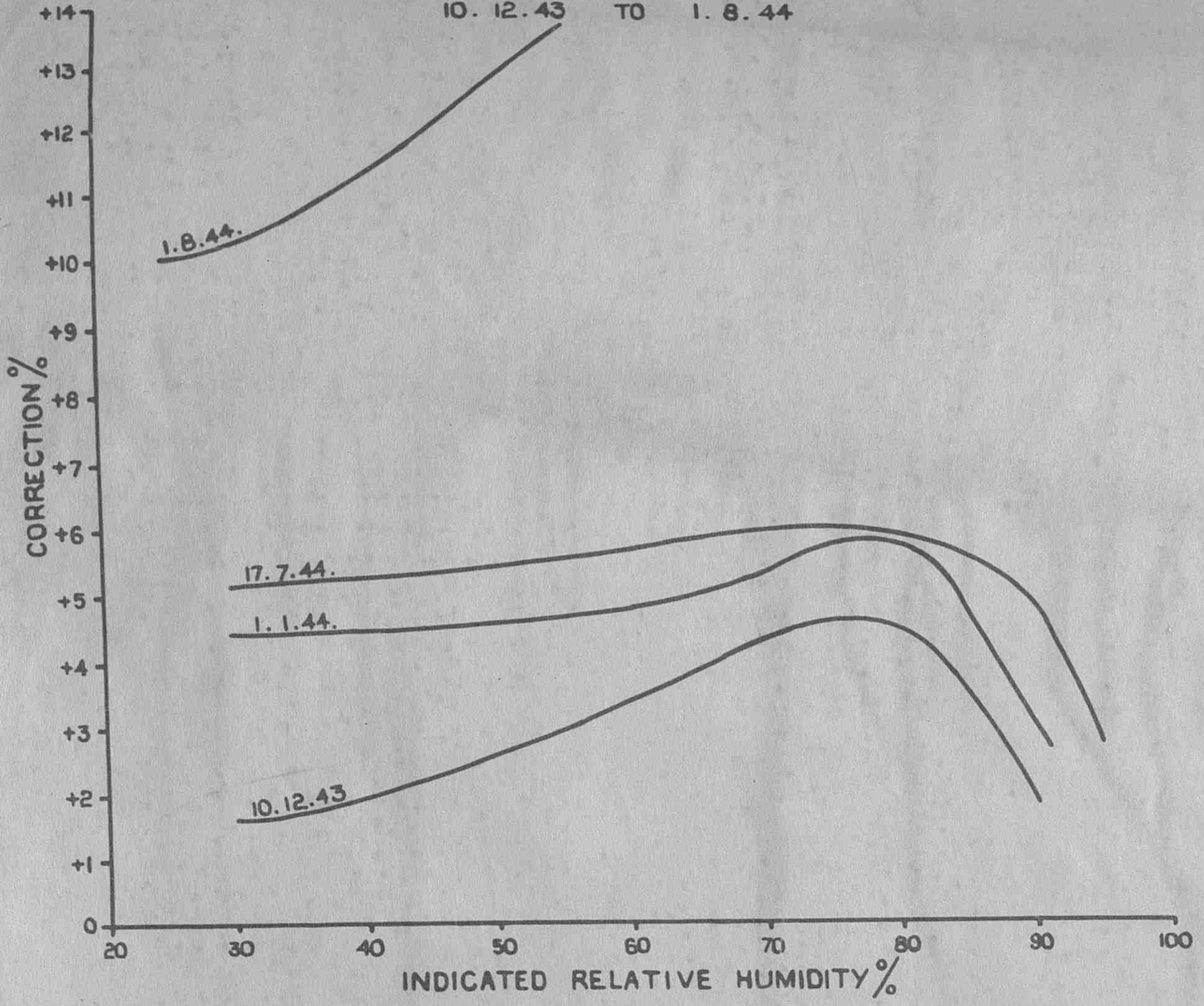
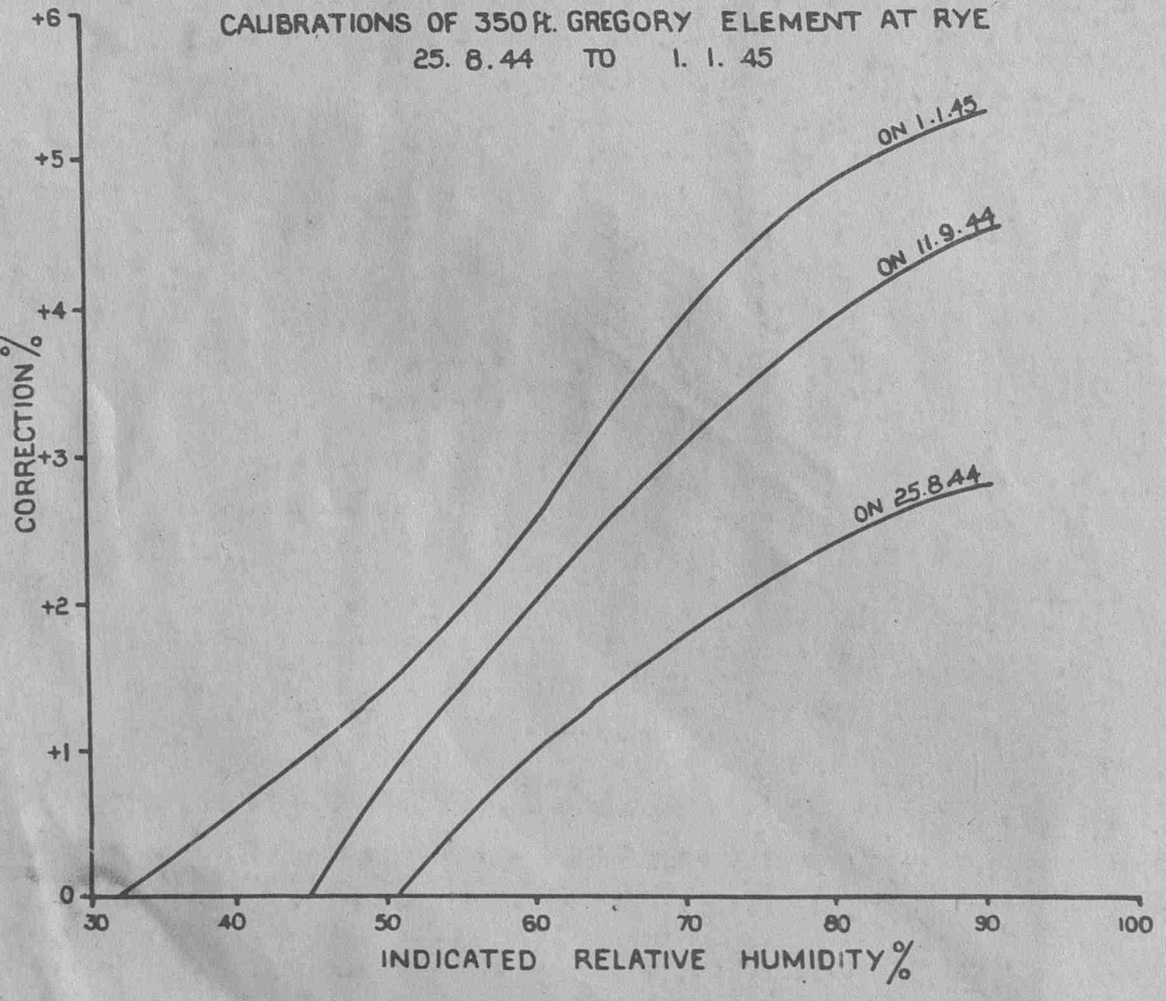


FIG. 15

CALIBRATIONS OF 350 ft. GREGORY ELEMENT AT RYE
25. 8. 44 TO 1. 1. 45



~~RESTRICTED~~

AIR MINISTRY

MRP 228
22nd March, 1945.

METEOROLOGICAL RESEARCH COMMITTEE

Classification cancelled (or changed to) unclassified
by authority of MRP 305 22 Oct 46
by John [illegible] date 22 Apr 57

Maximum and Minimum Temperatures in the Upper Air.

by

F.M. Jones.

AIR WEATHER SERVICE
TECHNICAL REFERENCE FILE

Since M.R.P. 150 was issued, an analysis has been made of upper air temperature records for some additional stations. The highest and lowest temperatures recorded in July and August, and in January and February have been tabulated for the following stations (Tables 1, 2, 3 and 4):-

(a) Abisko, Sweden - ballons sondes from 1921 - 29.[1]

(b) Agra, India - ballons sondes from 1928 - 36.[2]

(c) Batavia, Java - ballons sondes from 1910 - 15.[3]

(d) Munich, Germany - ballons sondes from 1906 - 14, 1922 - 28.[4]

The maximum and minimum temperatures for the year are also shown for Alaska, U.S.A. (Northern and Southern States)[5,6], England[7,8], and for Abisko, Agra, Batavia and Munich, (Tables 5 and 6). The absolute range for the year is given in table 7, and cases of temperatures below -88°C (185°A) are given in Table 8.

In levels below about 8km. the absolute range of temperature (see table 7) is much lower at Batavia than at other stations; the next lowest range is at Agra. In the higher levels also the range in the tropics is less than in other parts of the globe, though, especially from about 12 to 16km., the difference is not nearly so marked as in the lower layers of the troposphere; indeed the absolute range at about 12 to 16km. is very much the same in the tropics as in England or Munich. The American stations, Agra and Abisko on the other hand all hang very much together in showing a high range of temperature above 12km.

Heights of Observations. The observations for England, Abisko and Agra are given for heights in gkm., while those for U.S.A., Batavia and Munich are for km.

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5. Washington Weather Bureau. Upper air maximum temperatures at standard levels over United States and Alaska. Technical Note, Washington, 1943

/6.

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7. London, Air Ministry, Meteorological Office, Observatories Year Book, 1926-37.

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Table 1. Maximum Temperatures in July and August.

Height km	Abisko °C	Munich °C	Agra °C	Batavia °C
1	13	20	30	22
2	7	14	22	16
3	4	9	16	11
4	-2	3	12	7
5	-7	- 2	7	0
6	-13	- 7	4	- 6
7	-22	-13	- 1	-11
8	-26	-19	- 8	-18
9	-33	-25	-15	-24
10	-39	-31	-23	-31
11	-40	-38	-28	-39
12	-36	-43	-36	-47
13	-36	-42	-45	-55
14	-35	-42	-55	-63
15	-43	-44	-63	-70
16	-41	-44	-69	-77
17	-40	-44	-67	-82
18	-45		-63	-71
19			-63	-69
20			-63	-65
21			-58	-64

Table 2. Minimum Temperatures in July and August.

Height km.	Abisko °C	Munich °C	Agra °C	Batavia °C
1	1	7	19	17
2	- 4	1	11	12
3	-11	- 6	11	6
4	-17	-10	6	0
5	-22	-16	1	- 4
6	-29	-23	- 5	-12
7	-37	-30	-11	-17
8	-44	-35	-17	-26
9	-51	-43	-25	-35
10	-52	-50	-33	-43
11	-55	-58	-46	-51
12	-61	-62	-51	-60
13	-58	-64	-62	-69
14	-49	-57	-75	-79
15	-48	-56	-79	-84
16	-48	-56	-85	-83
17	-48	-54	-86	-85
18	-47		-86	-87
19			-83	-70
20			-76	-67
21			-71	-65

Table 3. Maximum Temperatures in January and February.

Height km.	Abisko °C	Munich °C	Agra °C	Batavia °C
1	- 4	- 7	23	22
2	- 6	6	15	18
3	-11	0	10	11
4	-17	- 7	3	7
5	-20	-14	- 3	1
6	-27	-21	-11	- 5
7	-38	-29	-17	-11
8	-44	-35	-22	-18
9	-47	-42	-28	-22
10	-43	-48	-33	-29
11	-49	-46	-35	-37
12	-48	-47	-41	-44
13	-47	-46	-49	-52
14	-46	-45	-53	-61
15	-46	-46	-59	-69
16	-45	-47	-59	-76
17	-46	-47	-63	-83
18	-46		-63	-84
19	-47		-59	-79
20			-55	
21			-55	

Table 4. Minimum Temperatures in January and February.

Height km.	Abisko °C	Munich °C	Agra °C	Batavia °C
1	-14	-13	3	16
2	-19	-17	1	12
3	-26	-23	- 5	5
4	-32	-26	-12	0
5	-38	-32	-19	- 6
6	-43	-39	-23	-10
7	-50	-45	-32	-19
8	-56	-54	-41	-26
9	-60	-60	-50	-31
10	-69	-61	-54	-40
11	-75	-63	-59	-57
12	-69	-65	-65	-61
13	-65	-59	-70	-69
14	-67	-60	-77	-74
15	-69	-56	-77	-80
16	-70	-58	-80	-83
17	-71	-58	-79	-88
18	-72		-79	-89
19	-71		-81	-82
20			-80	
21			-77	

Table 5. Maximum Temperatures for the Year.

Height	Alaska	Northern States	Southern States	Abisko	England	Munich	Agra	Batavia
km.	°C	°C	°C	°C	°C	°C	°C	°C
1	24	37	40	19	22	22	35	25
2	17	29	33	13	16	18	26	20
3	12	23	24	~~9~~4	9	10	17	15
4	6	14	14	~~2~~-2	2	3	12	10
5	0	8	8	- ~~4~~ -7	- 4	- 2	7	4
6	- 6	3	0	-~~10~~-13	-10	- 7	4	- 1
7	-14	- 5	- 5	-22	-16	-13	- 1	- 5
8	-20	-12	-10	-26	-25	-19	- 8	-10
9	-29	-19	-16	-~~35~~33	-30	-25	-15	-16
10	-36	-25	-22	-~~41~~39	-38	-31	-23	-22
11	-35	-31	-30	-39	-38	-38	-28	-29
12	-35	-32	-37	-36	-42	-42	-36	-35
13	-33	-33	-43	-36	-~~43~~39	-42	-45	-46
14	-33	-33	-47	-35	-44	-42	-53	-52
15	-31	-34	-45	-39	-~~44~~41	-42	-57	-59
16	-30	-36	-51	-39	-44	-41	-58	-65
17				-40	-43	-41	-55	-71
18				-42			-58	-71
19				-45			-55	-69
20				-45			-55	-64
21							-49	-62

Table 6. Minimum Temperatures for the Year.

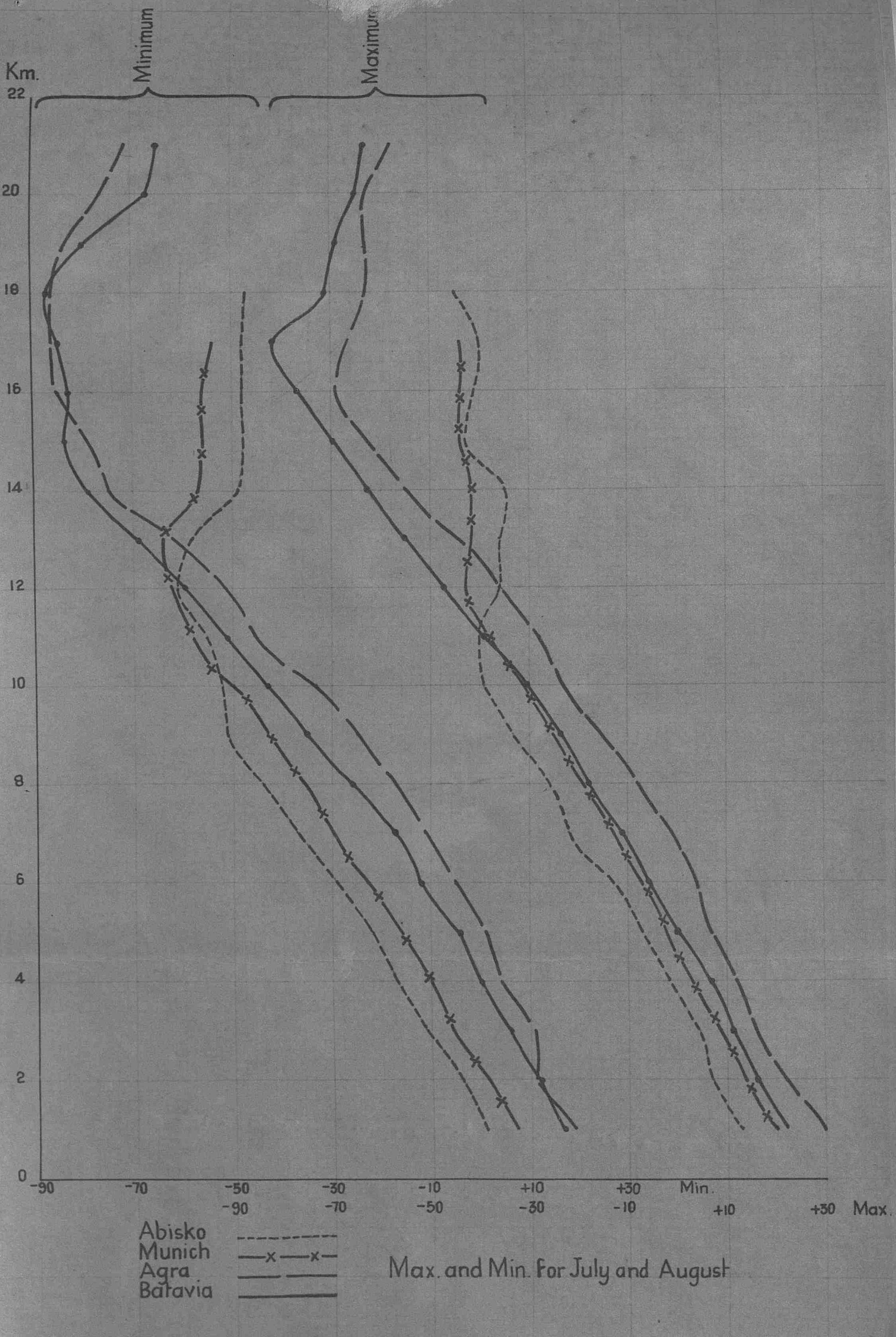
Height	Alaska	Northern States	Southern States	Abisko	England	Munich	Agra	Batavia
km.	°C	°C	°C	°C	°C	°C	°C	°C
1	-37	-36	-25	-19	-11	-13	0	16
2	-41	-37	-23	-23	-18	-17	- 1	12
3	-44	-43	-27	-31	-25	-24	- 6	5
4	-45	-46	-33	-40	-33	-29	-12	- 1
5	-50	-48	-38	-45	-38	-38	-19	- 9
6	-53	-51	-42	-53	-46	-45	-23	-15
7	-60	-56	-47	-56	-51	-52	-32	-20
8	-61	-60	-52	-62	-56	-55	-41	-27
9	-66	-66	-59	-64	-58	-60	-50	-36
10	-71	-71	-65	-69	-63	-63	-54	-43
11	-74	-74	-67	-75	-68	-66	-64	~~-52~~ ~~57~~ 82
12	-73	-77	-72	-77	-72	-68	-71	-61
13	-70	-81	-76	-69	-70	-72	-71	-69
14	-67	-78	-81	-70	-68	-67	-77	-79
15	-68	-80	-84	-70	-67	-68	-83	-84
16	-69	-79	-89	-70	-68	-65	-91	-89
17				-71	-70	-61	-90	-90
18				-72			-87	-90
19				-71			-87	-82
20				-71			-87	-78
21							-87	-76

Table 7. Absolute range of Temperature in °C. (Year)

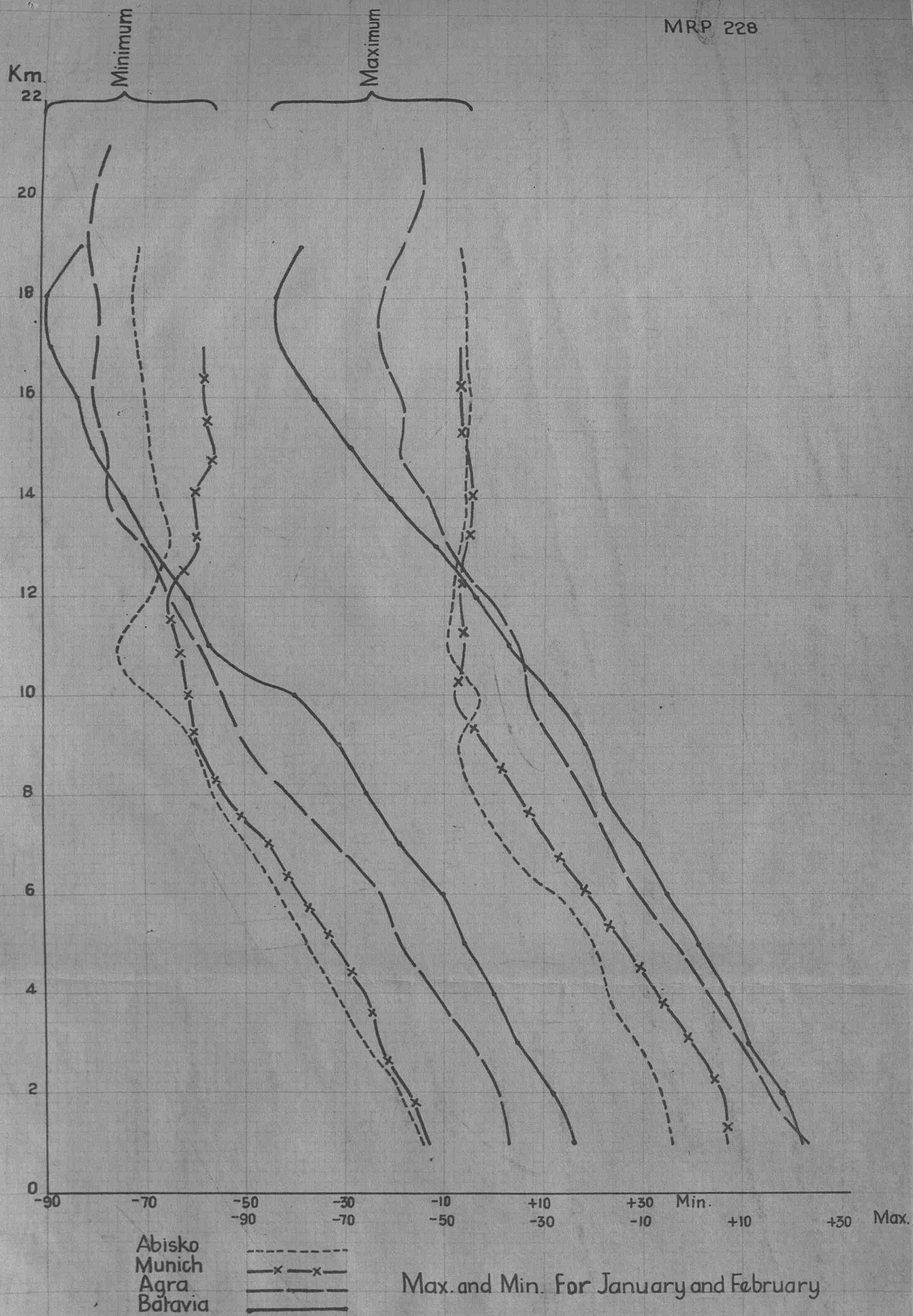
Height km.	Alaska	Northern States	Southern States	Abisko	England	Munich	Agra	Batavia
1	61	73	65	38	33	35	35	9
2	58	66	56	36	34	35	27	8
3	56	66	51	35	34	34	23	10
4	51	60	47	38	35	32	24	11
5	50	56	46	38	34	36	26	13
6	47	54	42	40	36	38	27	14
7	46	51	42	34	35	39	31	15
8	41	48	42	36	31	36	33	17
9	37	47	43	~~29~~ 31	28	35	35	20
10	35	46	43	~~28~~ 30	25	32	31	21
11	39	43	37	36	30	28	36	~~23~~ ~~28~~ 23
12	38	45	35	41	30	26	35	26
13	37	48	33	33	~~27~~ 31	30	26	23
14	34	45	34	35	24	25	24	27
15	37	46	39	31	~~23~~ 26	26	26	25
16	39	43	38	31	24	24	33	24
17				31	27	20	35	19
18				30			29	19
19				26			32	13
20				26			32	14
21							38	14
Mean 1-8 km.	51	59	49	37	34	36	28	12
Mean 12km. and over	37	45	36	32	26	25	31	20

Table 8. Cases of temperature below -88°C (185°A).

Temperature		Height	Date	Place
°C	°A	km.		
-88.5	184.5	16	July	San Antonia, Texas
-88.1	184.9	16	2. 5.12	Batavia
-90.9	182.1	16.5	4.12.13	Batavia
-89.3	183.7	17.7	5. 3.14	Batavia
-89.1	183.9	17.9	27. 1.15	Batavia
-89.7	183.3	17.5	8. 2.15	Batavia
-90.0	183.0	17.8	2. 5.15	Batavia
-89.0	184.0	17	17.11.36	Madras
-90.0	183.0	18	24. 4.32	Poona
-88.5	184.5	17	8. 3.30	Poona
-90.0	183.0	17	14. 5.30	Poona
-88.5	184.5	17	16.10.31	Agra
-91.0	182.0	16	4.10.28	Agra



Max. and Min. for July and August



Max. and Min. For January and February

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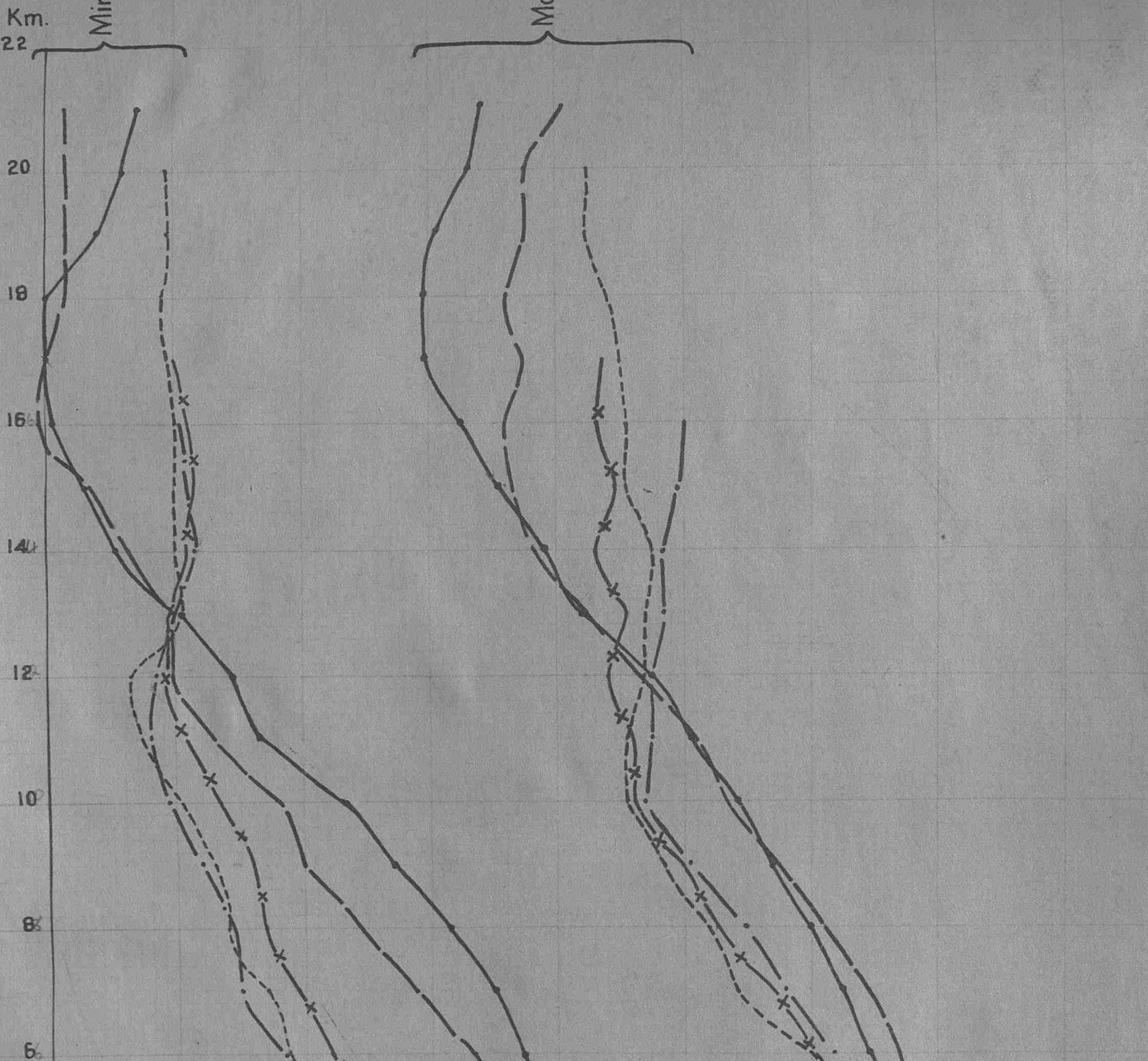
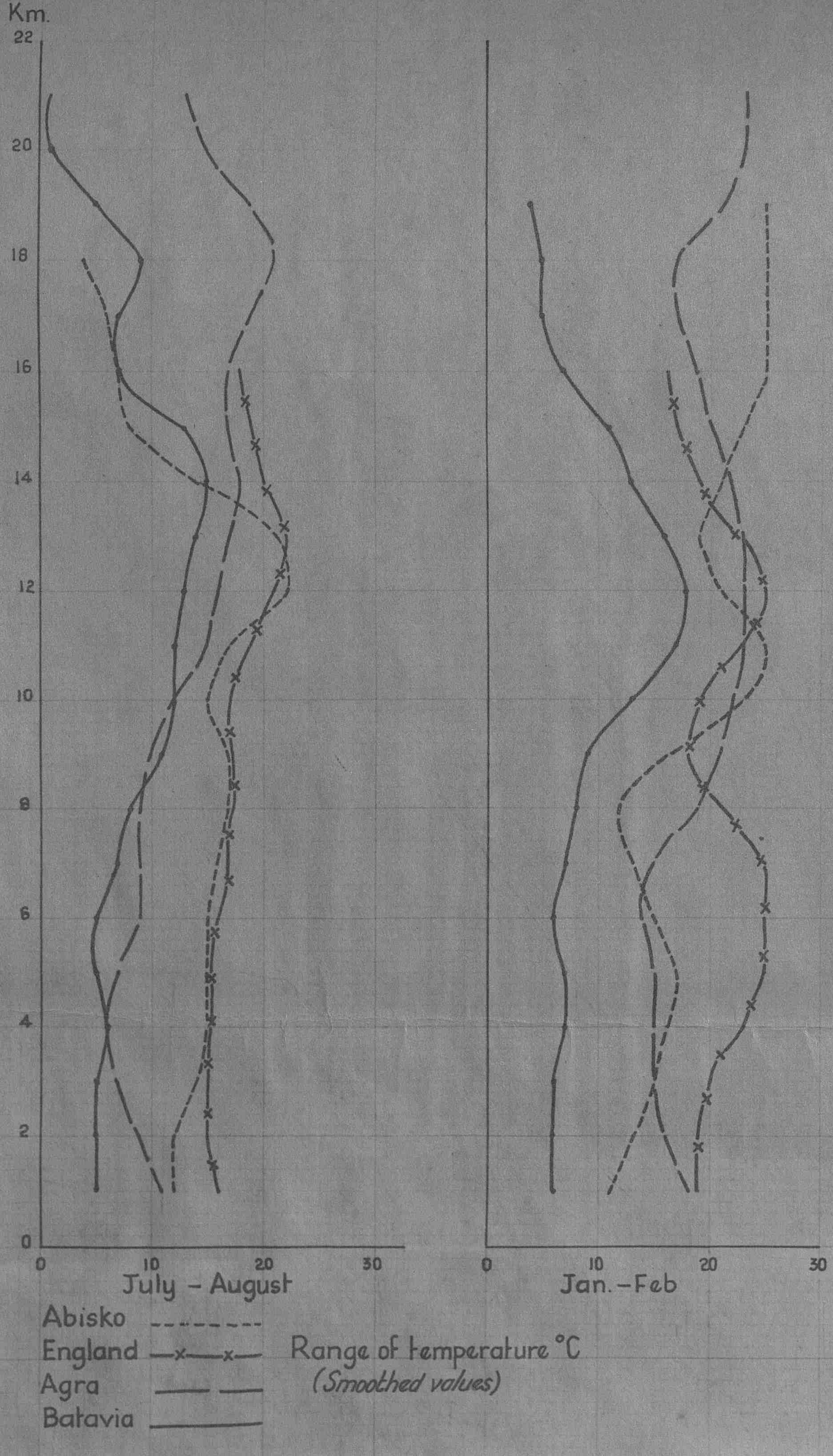


Max. and Min. Temperatures °C Year.



Range of temperature °C
(Smoothed values)

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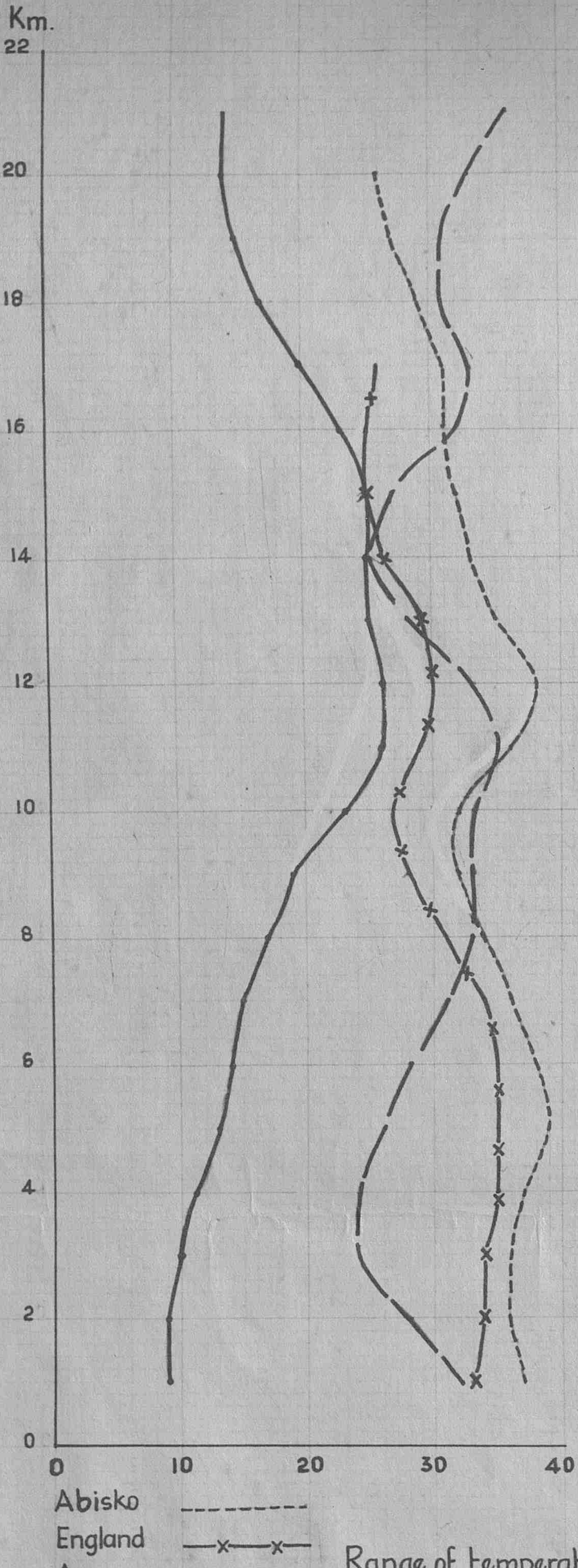

Range of temperature °C (Year)
(Smoothed values)

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MRP 233

MRP 305 12 Oct '46 METEOROLOGICAL RESEARCH COMMITTEE

9th April, 1945.

Development of a Rocket-borne Radio Sonde

by

N.K. Johnson.

1. Introduction.

The progress which has been made recently in the development of rocket projectiles raises the question whether this method could be usefully employed for lifting radio sonde instruments. The following paragraphs discuss the practicability of this suggestion from the technical point of view and compare the advantages and limitations of the method with the existing practice of employing a balloon.

2. Technical problems of development.

In order to ascertain whether the proposal is possible from the technical point of view, a visit was paid to the Projectile Development Establishment, and a record of the discussions which took place is given in the appendix to this paper.

On the assumption that the initial work should be restricted to employing a 3" rocket capable of reaching a height of about 30,000 ft., it will be seen that no insuperable difficulty is foreseen. In particular, the accelerations are not greater than that which a radio sonde can be made to withstand. The existing design of radio sonde would need considerable modification to withstand the accelerations, and the components would have to be re-arranged both to permit their accommodation within the 3" diameter of the rocket and to ensure their correct functioning when falling instead of rising. A number of minor technical problems would also require solving. The collaboration of the Projectile Development Establishment could be reckoned on in this development work.

3. Comparison of rocket and balloon.

The advantages and limitations of rockets and balloons as methods of lifting a radio sonde are considered below under various headings:-

(a) Height attainable.

The standard 3" rocket will attain a height of about 30,000 ft. as compared with about 55,000 ft. normally reached by a balloon. Whilst a height of 30,000 ft. would be useful for many routine purposes, if greater heights were desired it would be necessary to develop a larger size of rocket.

(b) Ease of operation.

The outstanding advantage of the rocket is its great simplicity of operation. This advantage would be of great value on board ship or under any conditions where it was desirable to avoid the filling and handling of a balloon.

(c) Weight.

The weight of a 3" rocket would be about 56 lbs. If hydrogen for filling a balloon is provided in steel cylinders the weight of cylinder required is about 60 lbs. The generation of hydrogen from special alloys would reduce the weight to appreciably below 50 lbs, although it is not yet possible to say whether this method of hydrogen generation will be practical and economical on a large scale. On balance, there would appear to be little to

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choose between rockets and balloons from the point of view of weight.

(d) Cost.

The cost of manufacture of a 3" rocket is about £5 which is substantially greater than the cost of a balloon and its hydrogen. On the other hand, it is understood that large stocks of rockets are likely to be available after the war.

(e) Signal reception.

In the case of a balloon ascent, the transmitter starts at the ground and its height increases with its horizontal distance from the recording station. On the other hand, a rocket-borne transmitter will start nearly overhead and its height will diminish as its horizontal distance increases. The N.P.L. have been consulted on the effect which these changes would have upon the strength of signal received at the recording station.

It appears that no serious difficulty would be experienced in the initial stage provided the rocket were fired 20 degrees out of the vertical. During the later stage, a transmitter falling from 30,000 ft. should normally remain audible down to 1,000 ft., by which time its distance from the launching point will not usually exceed about 25 miles.

4. Summary.

A rocket-borne radio sonde would provide a self-contained store which could be brought into use immediately and without the delay and trouble of filling and launching a balloon. These advantages would render it particularly suitable for use by Merchant Navy Reporting Ships and by Mobile Units in the field in wartime. It would also be valuable at places where the delivery of heavy equipment, such as hydrogen cylinders or hydrogen generators, presented difficulty.

There appears to be no insuperable difficulty in adapting the Kew Mark I radio sonde transmitter for use as a rocket-borne instrument, and it is considered that the value of such an instrument justifies the development work which would be involved.

APPENDIX TO MEMORANDUM

on

Development of a Rocket-borne Radio Sonde.

(Note on a visit to the Projectile Development Establishment).

1. With the object of examining the practicability of developing a rocket-borne radio sonde, a visit was paid by representatives of the Meteorological Office to the Projectile Development Establishment at Aberporth.

2. The following information was obtained regarding the characteristics of rockets:-

(i) P.D.E. suggested that the 3" rocket will probably be capable of carrying a radio sonde unit to a height of about 30,000 ft. It was agreed that if it was decided to proceed with the idea of developing a rocket radio sonde, then the 3" rocket should be used as a basis for the initial work. From the performance of this equipment, it would then be possible to design equipment capable of reaching a somewhat greater height should this be considered necessary.

(ii) The possibility of developing a rocket capable of reaching heights of the order of 200,000 ft. was mentioned, but it was agreed that this should await the results of work on the 3" rocket.

(iii) The 3" rocket consists essentially of two portions, the tail and the head. The former constitutes the propelling unit, and consists of a steel tube some 30" long and 3" in diameter, containing $12\frac{1}{2}$lbs. of cordite. The products of combustion of the cordite emerge from a Venturi shaped orifice at the lower end of the tail. Burning of the cordite occupies only about $1\frac{1}{2}$ seconds, after which the rocket travels as a free projectile. The total weight of the tail unit is about 48 lbs. The head is normally shorter than the tail unit and is pointed like a shell in order to reduce the air resistance. The weight of the head is also usually considerably less than that of the tail.

The following figures refer to the 3" rocket:-

Acceleration: Max.	-	90g.
Mean	-	25 - 30g.
Max. velocity:-	-	1600 - 1700 ft./sec.
Time to vertex: (30,000 ft.).	=	40 secs.
Weight of rocket:		
Tail	-	48 lbs.
Head	-	say, 8 lbs.
Cost.	-	about, £5.
Internal dimensions of head:		
Diameter	-	$3\,{}^{1}/_{16}$ ins.
Length		up to 2 feet.

(iv) The question of the possible heating of the head due either to conduction from the tail unit or to compression of the air over the head of the rocket was discussed. It was agreed that both these possible sources of heating would need to be borne in mind, but no difficulty was anticipated in introducing suitable non-conduction arrangements to reduce the effect to small proportions.

(v) A parachute would have to be provided to reduce the velocity of descent of the tail unit. A velocity of 60 ft. per second was regarded as permissible for this. To reduce it to 20 ft. per second would involve an impracticably large parachute. It would therefore be necessary to provide two parachutes, one for the tail unit and one for the radio sonde unit. This means ejecting the radio sonde unit from the rocket; it was considered that the acceleration involved in this process need not exceed the acceleration produced in launching the rocket.

(vi) The Type K rocket ejects a flare container whilst travelling at a speed of about 400 ft. per second. In order to reduce the speed of the container before its parachute opens, the centre of gravity is arranged to come at a different place from the centre of pressure, thus causing the container to rotate lengthwise. It would be necessary to avoid this occurring in the case of a radio sonde, since it would be difficult to design an instrument to withstand the accelerational forces produced by the rotation. This result could be achieved by arranging for the ejection to take place close to the vertex and also by making the centre of gravity coincide with the centre of pressure of the ejected container.

(vii) It was agreed that it would be necessary to provide the redesigned radio sonde with non-spillable batteries. No decision was reached on the best way of meeting this, and the problem was left over for the Meteorological Office.

(viii) To avoid sudden changes of pressure upon the radio sonde, it would be necessary to arrange that the holes leading to the inside of the head were located at the position of zero pressure on the outside of the rocket.

(ix) With regard to the employment of rockets in the tropics, it was stated by P.D.E. that the actual functioning was satisfactory up to a temperature of 120°F. Above that temperature the cordite propulsion was liable to burn with explosive violence.

As regards keeping properties in the tropics, P.D.E. stated storage conditions standardised for cordite would ensure the rockets remaining satisfactory for some years. The problem of preventing adverse affects due to moisture was already being tackled in connection with the problem of using rocket weapons in the Far East.

(x) A very simple form of projector could be used for launching the 3" rocket vertically. It would consist merely of 3 or 4 rods of about 1" diameter held between base and head plates. It would weigh about ½ cwt. and would cost only a few pounds.

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METEOROLOGICAL RESEARCH COMMITTEE

MRP 234
11th April 1945.

Note on the variations of Newfoundland ice.

by G. T. Walker.

Summary.

In the previous note on this subject it was pointed out that much of the Newfoundland ice came from regions to the northwest: also the main controls of ice are winds. Now for west winds to produce an abundant amount we want a pressure distribution high to the south and low to the north; and the only station on the north coast is Barrow (Alaska). We find that pressure there for October and November has a coefficient -.70 with the amount of ice in the following March to July, and by use of other data a somewhat better formula could be found. In spite of the great distance of Alaska from Labrador Fig. 1 shows considerable ground for accepting the surprising relationship; it is not until 1943 that war conditions appear to have made the Barrow observations unreliable. It may be that ten more years of trustworthy readings may confirm this mode of forecasting; but it would be unwise to confide in it as yet.

In my last note the relationships of Newfoundland ice with conditions over a wide area were examined; and in its concluding sentence attention was drawn to the belief, for which there is considerable evidence, that much of the ice in question comes from the arctic regions to the north-west of Labrador, regions for which no long series of records are available.

When observations are derived from isolated stations at great distances from headquarters, whether in hot or cold countries, we cannot be sure of complete reliability unless some check has been applied. But for the stations used in this note the inter-relations of the data are extremely consistent. The chief statistical obstacle is the shortness of the series of observations.

2. The only pressure series available here from the north coast of North America is one from 1921 to 1942 at Barrow in Alaska. About 500 miles to the south of it there is running west and east a line of observatories from Nome on the Bering Strait through Tanana and Eagle in Alaska to Dawson just over the Canadian border. From Nome to Dawson is about 800 miles, with Tanana midway between and Eagle close to Dawson.

Since winds are the most important controls of ice movement we should expect that an excess of Newfoundland ice would be preceded by northwest winds in the Canadian archipelago and, possibly, west winds along the north Alaskan coast; and we might hope to find that at such times pressure was lower on that coast than in the area to the south. For the relationship between the quantity of Newfoundland ice and the preceding monthly pressures at Barrow we find for the periods '21-'30, '31-'38,9 and '21-'38,9 the coefficients of the following table:-

Table 1.

	Aug.	Sep.	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.
'21-'31	-10	-32	-50	-64	-24	00	08	-24
'31-'38,9	+02	+06	-42	-74	00	20	-48	-18
'21-'38,9	-04	-14	-48	-68	-14	06	-22	-20

The control exercised by Barrow is at its height in October and November, roughly seven months before the ice reaches Newfoundland. With the O,N pressure (that of October, November) the relationships with the ice during the three periods just mentioned are all -.70. See Fig. 1.

3. In order to throw light on the pressure gradient we must examine the data of the inland stations. The coefficient of the ice with the preceding O,N pressure during the period '22-'31 at Nome, Tanana, Eagle and Dawson are -.22, -.16, .00 and -.12, values much less than the coefficient with Barrow. The difference is certainly not due to badness in the observations: for the D - F (December to February) pressure of Eagle has coefficients of .96 (13) and .98 (13) with those of Dawson and Tanana, while between Dawson and Tanana the coefficient is .96 (14): the numbers in brackets are the numbers of years in the data used. Such consistency shows accuracy of a high order.

The coast stations of Nome and Barrow are subject to other influences and the D - F relationship between them is .56 (9); for O,N it is .48 (10). With the previous three stations the coefficients are

	Tanana	Eagle	Dawson
Nome	.88(14)	.76(13)	.78(23)
Barrow	.44(9)	.38(9)	.36(9)

A further check lies in the relationships with the N. Pacific Oscillation (Memoirs R. Metl. Soc. IV, No.36, 57, 8: 1932) which plays in the Pacific Ocean a part similar to that of the N. Atlantic Oscillation in the Atlantic. The coefficients of the N. Pacific Oscillation with the inland winter (D-F) pressures are:-

Tanana	Eagle	Dawson
-.86(14)	-.92(13)	-.86(29)

and with Nome and Barrow -.86(22) and -.24(10).

4. It will be seen that while Nome has considerable association with the inland group Barrow shows appreciable independence in all cases. On the other hand unless the variations of Barrow differed from those of the stations to the south there would be no changes in the pressure gradient between them. So this independence does not involve unreliability. Moreover the Barrow pressure data are strongly supported by a comparison in Table 1 of the coefficients with Newfoundland ice during the two periods '21-'30 and '31-'39; there is also good agreement in October (-.50,-.42) and November (-.64, -.74), the months when there is a real relationship; and for the two months together we have -.70 and -.70. For the other months the differences are no greater than effects due to sampling will provide.

5. We may now return to the relationships of Newfoundland ice with October and November pressure in North America. For Nome we find a coefficient of -.20 (24), for Tanana -.08(14) for Eagle +.04 (14) and for Dawson -.12(26). Thus the negative relationship with the ice which is strong at Barrow is weak 500 miles to the south; but it is still -.32 at Juneau, 1100 miles southeast of Barrow. The coefficient between pressure there and at Barrow is +.34(20). Farther afield the coefficient of the ice with St. Louis pressure is .04 ('06-'21), -.40('22-'39) and -.24 ('06 -'39): between Barrow and St. Louis we have -.20 ('21-'30), + .23 ('31 -'40) and +.02 ('21-'40).

6. The immediate result of the analysis is a forecast at the beginning of December of the Newfoundland ice of March to July. That given solely by Barrow pressure of October and November has a coefficient R of .70. If we include also the effects of pressure at Nome and Eagle we find

(ice) = -.86 (Barrow pressure) -.33 (Nome pressure)
+.67 (Eagle pressure) (1)

where putting a quantity into brackets indicates that we take its "proportional departure" - its departure divided by its standard deviation. Then R between the forecast given by (1) and the actual ice is .82.

A further improvement could be made by using in addition data of the immediately succeeding December to February. A reference to Fig. 7 of the last paper[x] shows the temperatures at Bucharest and Athens as having coefficients of -.62 and -.58. But all these series must be extended before satisfactory conclusions can be reached.

7. Unhappily the monthly pressure data from Barrow suggest difficulties. In 1942 the values for two months are missing and in 1943 eight. In the last war there were gaps in the series of a number of countries and at such times reliability must diminish. At Barrow, as Fig. 1 shows, the pressure (reversed in the Figure) in 1944 is phenomenally high; and the data show values for October, November and December above normal by 0".41, 0".60 and 0".75, which are decidedly improbable.

8. On the physical side also there are difficulties. The ice which flows from the west into Baffin Bay and Davis Strait emerges through Jones Sound, Lancaster Sound or Hudson Strait, and perusal of literature by L.Mecking and E.H. Smith[x] leads to the expectation that ice would take something like 110 days to traverse the 1,200 miles from Lancaster Sound to the mouth of Hudson Strait, and about 80 days for the 1,000 miles thence to Newfoundland. So ice arriving off Newfoundland in May might have to leave Lancaster Sound sometime in November. From Barrow to Lancaster Sound is about 1,500 miles, and the low pressure at Barrow in October and November might have to produce its effect at a considerable distance to the east - probably somewhere in the archipelago - in order that the consequent increased flow of ice might reach Newfoundland in May. But the speed of the currents is not known exactly and there is uncertainty in these estimates.

It is clear that the Barrow pressure observations cannot be used as the foundation of a forecast until further years have established their accuracy and the closeness of their relationship with Newfoundland ice.

Mecking, L. Die Eisdrift aus Baffin Bay... Ver.d.Inst.f.Meereskunde 7, 1-132, Berlin 1906.

Smith, E.H. The Marion Expedition. U.S.Treasury Dept. Bulletin 19, Part 3. 1931.

[x] MRP 221.

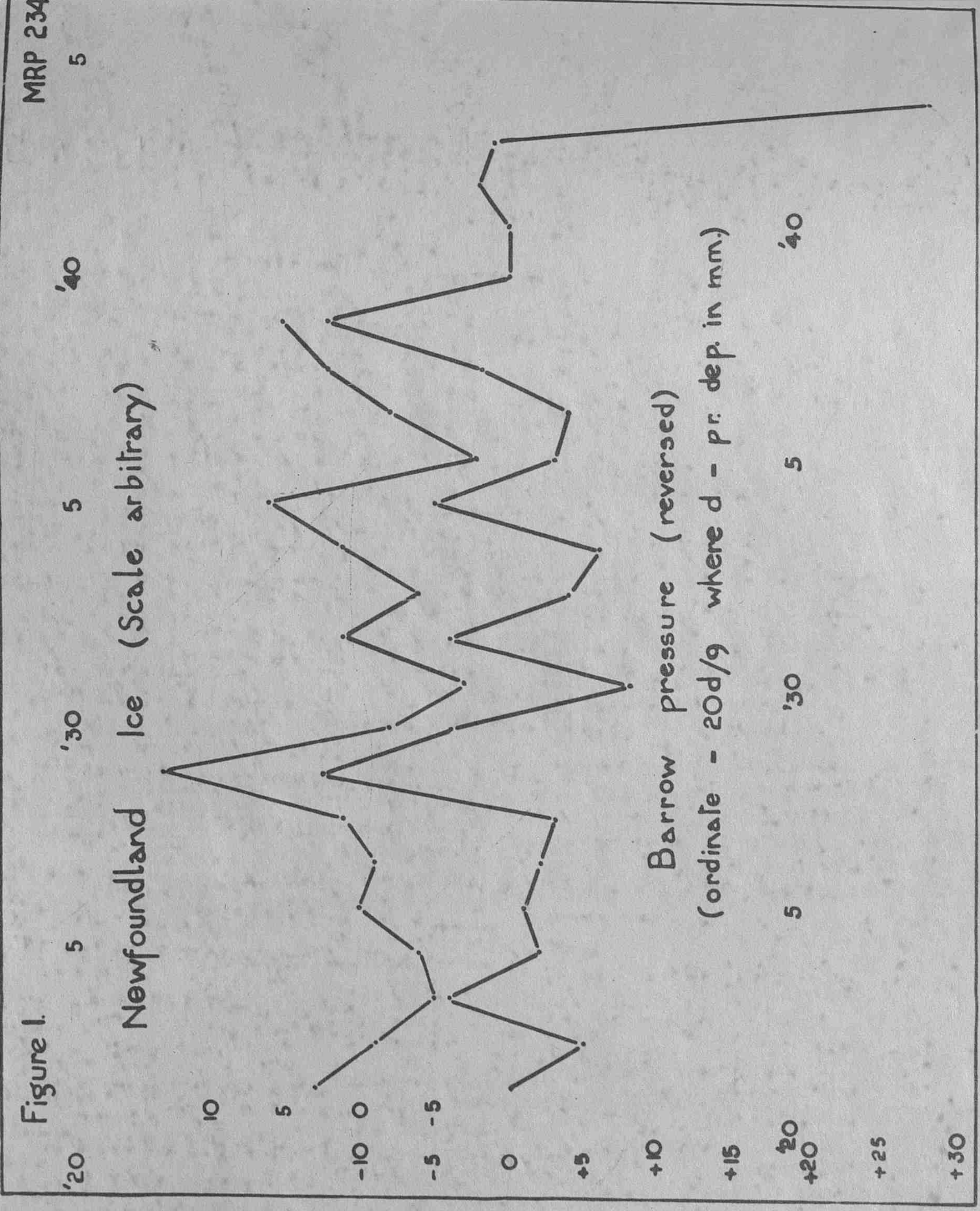

Figure 1.

2. To be added at end of paper:-

"Corrected pressure data recently received for Barrow for 1942 and 1943 remove the anomalies referred to in paragraph 7 and shown in Figure 1. The departures from normal for the last six years plotted in figure 1 were

-12, 0, 0,-2, -1, 28.

With the corrected values the last mentioned value is replaced by 1. Figure 1 should be amended accordingly".

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

Amendment to MRP 234.

Note on the Variations of Newfoundland Ice.

by

G.T. Walker.

1. To be inserted at the end of § 6.

"It may be, however, that the situation is not as favourable as these figures seem to imply. If we have two series between which the true correlation coefficient r' is .00 it is known from David's tables that the chance of a coefficient r of .7 arising from 18 years of data is .0006, or about 1 in 1670. If the true coefficient r' is .1 the chance of r = .7 is about 1 in 500, and so on as in the table:-

r'	.0	.1	.2	.3	.4	5.	.6	.7
P	.0006	.002	.006	.017	.047	.118	.270	.535
1/P	1670	500	165	59	21	8.5	3.7	1.9

So if we get .70 as the coefficient between the series the smallness of the likelihood of its arising from random numbers does not rightly lead to the belief that r' = .7. With a true coefficient r' of .5 the chance of getting r = .7 is 1/8.5; and if r' = .4 the change is 1/21. Thus a true value of .5 is reasonably possible and a change from .7 to .5 would materially reduce the value of a seasonal forecast".

2. To be added at end of paper:-

"Corrected pressure data recently received for Barrow for 1942 and 1943 remove the anomalies referred to in paragraph 7 and shown in Figure 1. The departures from normal for the last six years plotted in figure 1 were

-12, 0, 0,-2, -1, 28.

With the corrected values the last mentioned value is replaced by 1. Figure 1 should be amended accordingly".

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METEOROLOGICAL RESEARCH COMMITTEE.

The design of a Photoelectric Deposit Indicator for use with a Frost-point Hygrometer and notes on the functioning of Frost-point Hygrometers at low temperstures.

by

G.M.B. Dobson

and

B.M. Cwilong.

In MRP 212 we described a form of photoelectric deposit indicator for a frost-point hygrometer. While the arrangements of the illumination etc. there described worked well for air of frostpoints above about -60°C it was found that with drier air, giving a very fine deposit on the thimble, it was better to place the photoelectric cell normal to the thimble and the arrangement shown in Fig. 1 was adopted. This design has proved entirely satisfactory. Provided that the surface of the thimble is in an "active" condition (see below) frost-points of -80°C can be measured with ease, individual readings having an accuracy of ±0.5°C or less and each reading taking less than 30 seconds. This should be compared with an accuracy of about ±2°C with the eye observation instrument where each measurement takes about 5 minutes. There is no doubt that when very low frost-points have to be measured the disadvantages of the extra elaboration of the photoelectric instrument compared with the simple eye-observation instrument are many times outweighed by the advantages of greater ease in use and increased accuracy. The present design is regarded as being satisfactory for a standard instrument. The instrument is waiting trials in the air at Boscombe Down until the new Mosquitoes arrive there.

The principle of the instrument is seen from Fig. 1. Light is concentrated on to the thimble surface (in this case polished gold being found satisfactory,) and the directly reflected light passes into a light-trap where it is absorbed by multiple reflections from polished black surfaces. Light scattered by the deposit on the thimble is concentrated on to the photocell, stray light being cut out as much as possible. The jet supplying the air on to the thimble is a slot some 10 mm. long by 1 mm. wide and directs the air stream on to the thimble at an angle of about 10°. This produces a uniform deposit on the thimble over an area somewhat larger than the illumated spot. The jet is normal to the plane shown in the figure.

Since the deposit cannot be observed by eye there may be some difficulty at the higher temperatures in knowing whether the deposit is dew or hoar-frost. Provided that the surface of the thimble is not too clean it will probably be enough to cool the thimble to about -40°C so as to freeze the deposit, before making a measurement.

It will be noticed that this type of instrument is suitable for incorporating into an entirely automatic hygrometer where the temperature of the thimble is maintained always at the frost-point of the air supply. The fact that the actual amount of deposit on the thimble is not important - provided that it is not changing - should make the automatic control of the temperature easier. Further, since the thimble will be continuously cold, the deposit should always be in the form of ice and also the surface of the thimble may in these conditions be expected to remain in an "active" condition (see below.)

The following critical tests of the instrument have been made in the laboratory with satisfactory results:-

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/.(a)

(a) To see what effect the thickness of the deposit had on the observed frost-point.

Observations were made using a very thin, a moderate and a very thick deposit, the order of the observations being shown by the figures within brackets. The air supply was obtained by passing air over silica gel. It will be noticed that the air supply becomes slowly drier as the connecting pipes dry out. All the individual measurements are given so that the casual variations can be seen.

Observed Frost-points

Thin Deposit	Moderate Deposit	Thick Deposit
-	-57.2 (1)-57.1 Mean-57.2°C. -57.3	-58.6 (2) 57.8?Mean 58.4°C -58.6
-57.5 (3)-57.5 Mean-57.5°C -57.5	-58.2 (4)-58.0 Mean-58.1°C -58.0	-57.8 (5)-58.0 Mean -58.0°C -58.2
-58.0 (6)-58.1 Mean-58.1°C -58.3	-58.3 (7)-58.4 Mean-58.5°C -58.7	-58.5 (8)-58.3 Mean-58.5°C -58.7
-58.2 (9)-58.2 Mean-58.2°C -58.2	-	-
Means -57.9°C	-57.9°C	-58.3°C

Since the thick and thin deposits used on the above test were outside the range of thickness usually used, no appreciable change of observed frost-point with deposit-thickness is likely.

(b) To see if the brightness of the illuminating lamp caused any change in the observed frost-point (e.g. through heating of the deposit on the thimble.)

For this test the lamp was run alternately on 4 and 6 volts (the normal rating) while many measurements of the frost-point were made.

Mean observed frost-points.

Lamp on 6 volts	4 volts
-50.7°C.	-50.2°C

Since the energy radiated by the lamp is much greater when on 6 volts than on 4 volts (which was the minimum on which the instrument would work satisfactorily) it may be taken that there is no appreciable effect.

(c) To see if the rate of air flow through the jet makes any difference to the observed frost-point.

The pressure on the jet was adjusted alternately to 4 ins. of water and 24 ins. with the following results:-

.....

Observed frost-points

Pressure on jet					
24 ins.			4 ins.		
(1)	-54.8 -54.5 -54.8	Mean-54.7°C	(2)	-55.3 -55.2 -55.2	Mean-55.2°C
(3)	-56.2 -56.0 -56.0	Mean-56.1°C	(4)	-56.2 -56.0 -55.9	Mean-56.0°C
(5)	-56.2 -56.0 -56.0	Mean-56.1°C		-	
Mean -55.6°C			Mean -55.6°C		

(d) To test that the observed frost-points are really the true frost-points of the air supply.

It is not easy to get an air supply with an accurately known low frost-point. Air partially dried by silica gel was passed through a copper spiral and a cotton wool filter both cooled in a bath of cold acetone. It is necessary to have the filter to avoid snow being blown through into the instrument where it would evaporate. Unless the pipes have been very carefully dried they will give off water vapour, while the dried pipes and cotton wool when cooled to a low temperature may take up water vapour and give air with a frost-point below the temperature of the bath. In practice the air was passed for a long time under the same conditions when the observed frost-point became nearly steady. In this way the observed frost-points were found to agree with the bath temperatures to within ±0.5°C at -58°C and -68°C and to within ±1°C at -78°C. In all probability the differences between the observed frost-points and the bath temperatures represent real differences between the frost-points of the air and the bath temperatures.

The general conclusion from these tests is that under normal conditions of use in an aeroplane the error in the observed frost-points would be negligible compared with the variations which are observed from minute to minute even when flying level.

Several unexpected peculiarities have been observed while making these tests with the frost-point hygrometer at very low temperatures.

(1) As briefly reported in M.R.P. 212, when the thimble is cooled below the frost-point of the air supply the rate of deposition of ice increases as the thimble temperature falls until it is about 10° to 20°C below the frost-point of the air supply. As the thimble is cooled further the rate of deposition decreases and when the thimble is very cold (about -100°C) it becomes very slow indeed.

(2) If the true frost-point of the air supply be first observed, then the thimble be cooled to a very low temperature (about -120°C) and then warmed up again, it is always found that ice continues to be deposited on the thimble until it is some 5°C above the previously observed frost-point.

(3) Using air having a frost-point of about -80°C, when the thimble is first cooled little or no deposit is formed for some minutes after which the

deposit begins to form very slowly and after the thimble has been cooled for some time the rate of deposition is much faster as though the thimble surface had gradually become "activated" in some way. If the thimble is slightly warmed so that the deposit evaporates and then cooled again it is still "active" but if warmed up to room temperature for some time it looses this "activity".

We now consider these effects in turn in more detail:-

(1) In order to throw light on this effect experiments were made with the object of finding how the rate of deposition depends on

(a) the temperature of the thimble surface
(b) the absolute humidity of the air supply
(c) the velocity of the air jet playing on the thimble.

Since the amount of the deposit is in all cases very small it would be almost impossible to measure the rate of growth accurately but using the photoelectric deposit indicator described above we have taken its indications as a rough measure of the amount of deposit on the thimble, since the reading of the instrument is proportional to the photoelectric current produced by light which is scattered by the deposit. It is, however, known that the following errors will come in:-

(i) When the deposit is very thick a small additional deposit will cause a much smaller increase in the scattered light than if the same deposit were formed on a nearly clean thimble.

(ii) At very low temperatures the deposit first formed appears blue and since the photoelectric cell used is less sensitive to blue than to red light the photoelectric current will be smaller in these conditions.

(iii) Under the conditions when the deposit appears blue, the ice particles are small compared with the wavelength of light and will scatter light approximately in proportion to the square of their volume so that relatively little light will be scattered per unit mass of deposit in the early stages of deposition at very low temperatures.

Owing to the effect mentioned under (3) it is difficult to get accurate figures for the rate of growth of deposits but the values in Fig. 2 certainly show the effect correctly in a qualitative manner. The diagram shows the variations in the rate of growth of the deposit with thimble temperature, for air of three different absolute humidities, the jet velocity being the same throughout. The following are the main conclusions:-

(a) In all cases the rate of deposition reaches a maximum when the temperature of the thimble is some 10° to 20°C below the frost-point of the air supply and for lower thimble temperatures it falls off to a value which is roughly constant at still lower temperatures.

(b) The maximum rate of growth of the deposit is roughly proportional to the absolute humidity of the air supply. The same is true for the rate of deposition at very low thimble temperatures where rate is roughly independent of temperature.

(c) Other observations show that the rate of growth of the deposit changes very little with large changes in the velocity of the air jet playing on the thimble.

From the three observed curves shown in figure 2 it may be inferred that with an air supply having a still lower absolute humidity, a curve such as ABCD would be obtained. Under such conditions it would be very difficult to get a deposit at all since it is only for thimble temperatures between A and C that any appreciable deposit will form and even then the growth will be very slow. Further it must be remembered that when taking observations in the upper atmosphere one does not, as in the laboratory, know beforehand what frost-points to expect and it would be extremely easy to cool the thimble through the range ABC without seeing any deposit. This emphasises the caution already given that inferences are only trustworthy when there is a definite deposit on the thimble and when this is seen both to increase and to decrease with change in thimble temperature.

(2) It was at first thought that when the thimble was cooled to a very low temperature the Bakelite wall of the chamber which was in contact with the thimble would be cooled below the frost-point of the air and ice would be deposited close around the thimble. This ice would evaporate when the thimble warmed up and would increase the humidity of the air in the chamber. Though the jet of fresh air is continually playing on the thimble, some of the relatively damp air might get mixed with the air of the jet and thus temporarily raise its humidity. A special thimble was therefore made as shown in Fig. 3. With this arrangement it seems almost impossible for air from the chamber to be mixed with the air in the jet before it passes over the thimble. Further the side walls being part of the copper thimble will be at its temperature while the glass cover to the thimble being separated from the thimble by a stream of relatively warm air should keep well above the frost-point and have no ice deposited on it. However the effect of the abnormally high apparent frost-point immediately after cooling the thimble to a very low temperature was shown with this type of thimble just as much as with the open thimble previously used. It is thus very unlikely that the effect is due to the evaporation of ice formed when the thimble is very cold.

If the thimble be cooled as above to a very low temperature and then warmed and held constant at a temperature about 3°C above the true frost-point of the air supply, ice continues to be deposited for about a minute, deposition then stops and soon after the whole deposit rapidly evaporates after which normal frost-points are found again. One might suppose that ice of a different form and having a somewhat different vapour pressure was formed at very low temperatures and that this slowly changed back to normal ice at higher temperatures. Against this view is the fact that the effect was very similar with air supplies of frost-points -58°C, -68°C and -78°C. There appears to be no reason why the temperature at which the supposed abnormal type of ice changes back to normal ice should depend on the frost-point of the air supply.

Fortunately this effect is unlikely to occur during the normal working of the hygrometer.

(3) This effect is of much practical importance in making measurements on very dry air. When first using the instrument with very dry air it is frequently impossible to get any reading since no deposit occurs at any temperature. If the thimble is kept cool for a few minutes a slight deposit slowly begins to grow and the rate of growth increases until accurate observations can easily be obtained. It appears that the surface of the thimble has in some way become "activated". The "activity" does not disappear if the thimble is warmed enough to evaporate the deposit but does disappear if it is warmed to room temperature for some time. It is therefore not due to the gradual growth of small ice crystals which act as nuclei for further sublimation. The observed frost point of the air supply is the same whether the thimble be in a fully "active" condition or only partly "active". The "activity" is very variable and the rate of deposition of ice is seldom constant in two successive tests where the conditions appear to be alike. This effect makes it impossible to draw curves such as those of fig. 2 accurately since the values do not repeat themselves. (The very small rate of deposition at very low temperatures is, however, always shown and there is no doubt about the correctness of the general shape of the curves though it is doubtful whether the curves should ever cross one another.)

It was thought that the "activity" might be due to fine silica gel dust blown over from the drying vessel and deposited on the thimble by thermal precipitation when cold but blown away again when it was warm. However inserting a cooled cotton wool filter in the air stream produced no difference. So far, in spite of many tests, we have been unable to find what it is due to or what conditions specially favour a very "active" state of the thimble. Fortunately when the instrument is in continuous use the thimble is generally in an "active" condition and favourable for measurements at very low temperatures.

ADDENDUM.

While there are still some difficulties to be explained, the two following experiments seem to indicate that the absence of any apparent deposit of ice at very low thimble temperatures is due to the deposit being in the form of a glass which does not scatter light.

(1) The hygrometer with photoelectric deposit indicator was well dried by warming while a stream of dry air was passed through it. A supply of very dry air was obtained by drying air to a frost-point of about -70°C with silica gel and then passing it through a copper spiral cooled in liquid air. The special type of thimble described above was used in which only fresh air direct from the air supply can reach the thimble. The procedure was as follows:-

(a) Very dry air was passed through the instrument.
(b) The thimble was cooled to -120°C, there being of course no deposit.
(c) Air of frost-point about -50°C was passed through for a few seconds when a visible deposit formed on the thimble and on the glass covering the slot in the thimble.
(d) Dry air was again passed and the thimble temperature still held at about -120°C until all visible deposit had evaporated from the glass cover plate - that on the cold thimble of course remained.
(e) The thimble was allowed to warm up when the visible deposit greatly increased as the temperature passed about -70°C and finally evaporated again. During this warming the very dry air supply was passing.

It seems that this apparent increase in the amount of deposit is most probably due to a glassy deposit formed at the low temperature changing to the crystaline form. The water could not have come from the air stream since very dry air was used. It could not have come from the walls of the vessel since the design of thimble was such as to prevent this. It is very unlikely that it could have come from the glass cover-plate since all visible deposit had evaporated and in any case water vapour from the glass plate would be very rapidly removed by the air stream and would have little chance of coming in contact with the thimble.

(2) In an experiment made in a vacuum, a molecular beam of water vapour was directed on to a metal surface cooled in liquid air. Here a glassy layer of ice was certainly formed since interference bands were shown. When the cold metal surface was warmed up the glassy layer became opaque. The existance of glassy ice has, of course, often been shown before.

A difficulty is to understand why a fine deposit of small ice crystals does not "activate" the surface of the thimble at low temperatures. The same phenomenon of the apparent slowing down in the rate of deposition of ice at very low temperatures occurs whether the surface of the thimble appears clean or covered with a thin layer of ice crystals. What could "activate" the surface better than a layer of small ice crystals?

Fig. 2

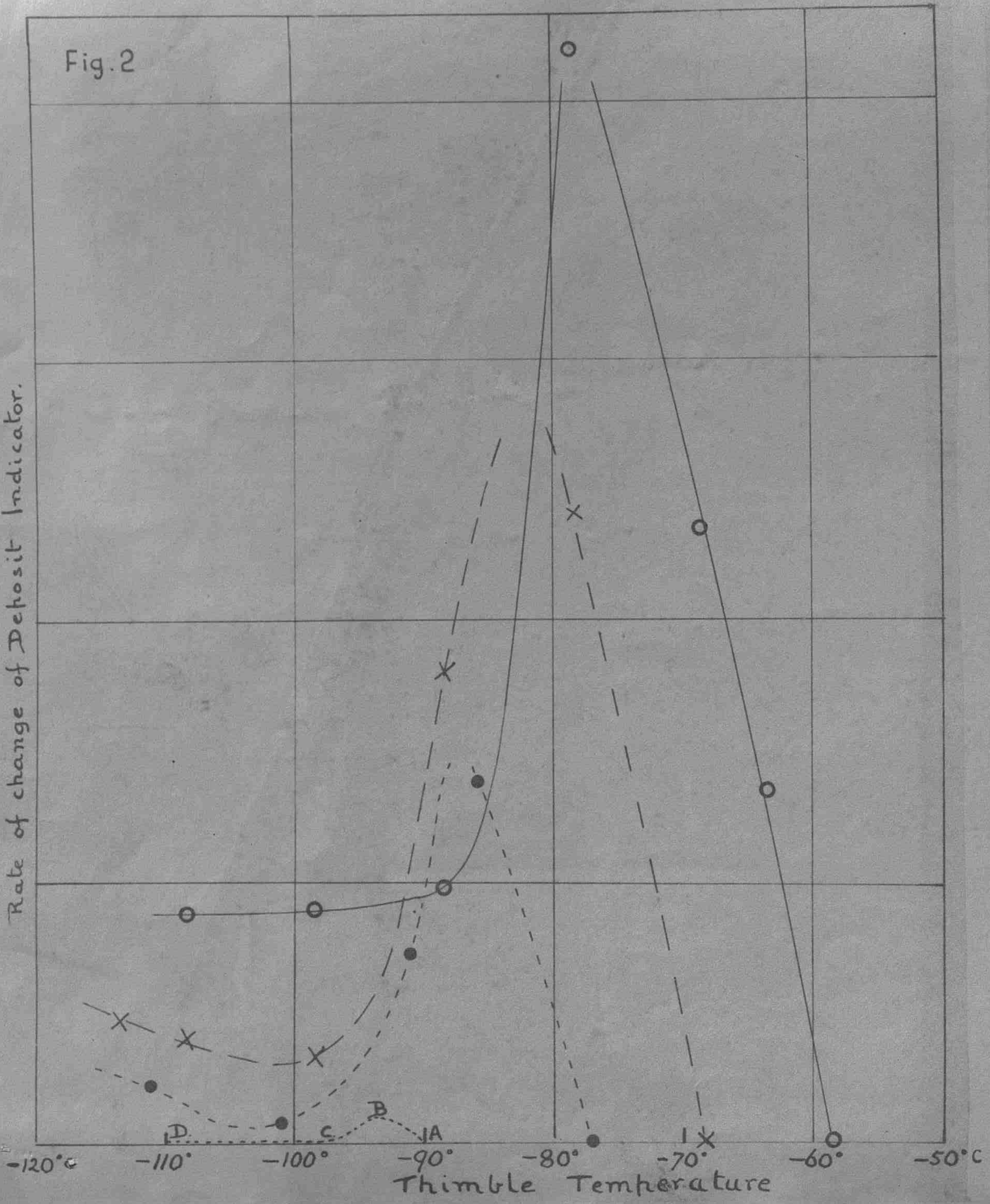


Fig. 1

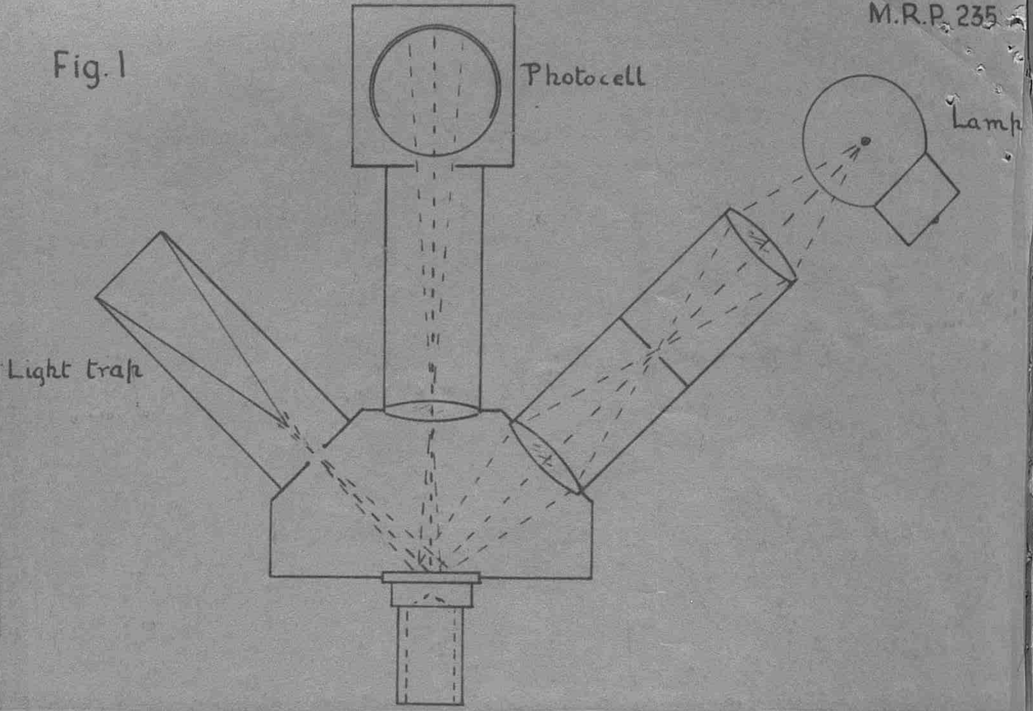

Photocell
Lamp
Light trap

Fig. 3

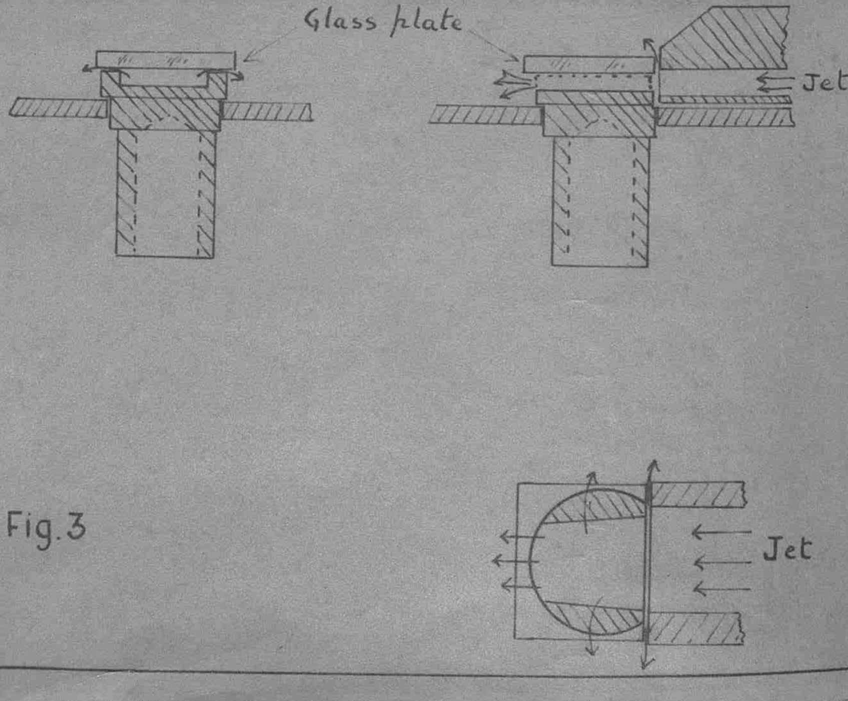
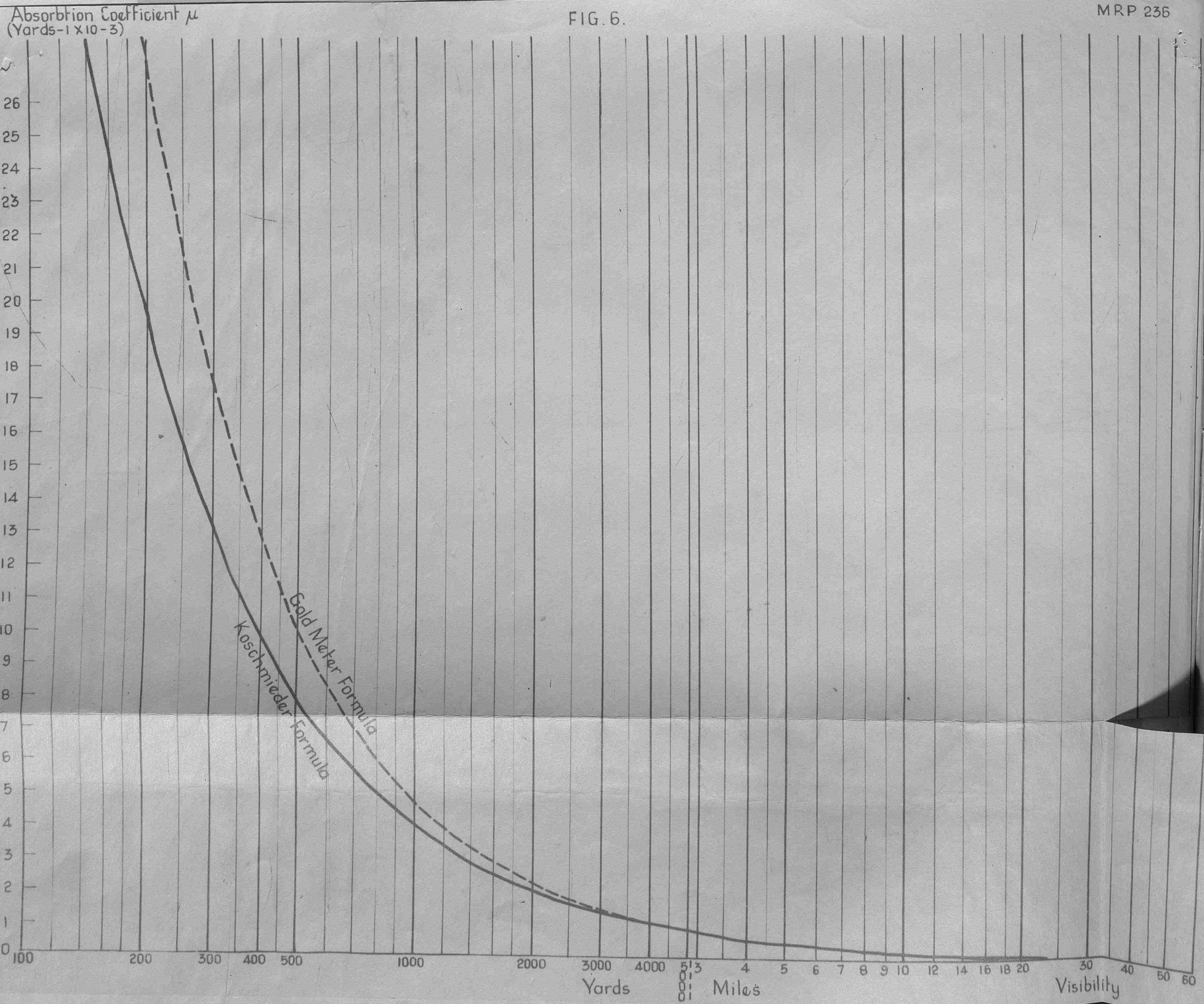

Glass plate
Jet
Jet

FIG. 6.


Absorbtion Coefficient μ
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Gold Meter Formula
Koschmieder Formula
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AIR MINISTRY

MRP 236
21st April 1945.

METEOROLOGICAL RESEARCH COMMITTEE

REPORT ON FIELD TRIALS OF THE PHOTO-ELECTRIC VISIBILITY METER

by

J.R. Bibby.

Classification cancelled (or changed to) Unclassified
by authority of M.R.P. 305 12 Oct. 1946
by date 16 Apr 59

1. Introduction.

This report describes a trial of an instrument developed by Dr. W. R. G. Atkins, described in MRP 157. The trial was carried out at R.A.F. Station, Defford, Worcs.

2. Experimental Arrangements.

The apparatus used was made up by the Instruments Branch of the Meteorological Office to the design of Dr. W.R.G. Atkins. Some modifications, as indicated below, were introduced to reduce the amount of stray light reaching the photocell.

Fig. 1 shows the photo-electric cell PC (of the selenium rectifier type) mounted in a box with a lens L in front. A screen S across the box, with a hole H (8 mms in diameter) at the focus of the lens, cuts off all light except that coming from the direction of the source of light being used. A small shutter SH covering the hole in the screen could be remotely controlled by means of an electro-magnet. The photocell was connected to a Tinsley mirror galvanometer which gave a full scale deflection (100 divisions) for a current of about 4 microamperes. The photocell box was mounted on the outside wall of the office, 7 feet above ground level, while the galvanometer and the switch controlling the shutter were located inside the office.

The source of light was a "portable cloud searchlight", consisting of a 12 volt 36 watt bulb at the focus of a parabolic mirror 7½ inches in diameter. This was mounted 6½ feet above ground level on a rigid post 334 yards from the photocell. Current for the lamp was obtained from a 12 volt accumulator nearby, the lamp being switched on and off by a relay remotely controlled from the office. The voltage of the battery could be measured across the switch (in the office) which operated the relay, but only when the switch was off.

The searchlight beam was very narrow, as shown by Fig. 2 which gives the intensities at various points on the cross section of the beam (334 yards from the light.) The figures are galvanometer readings with the photocell in the various positions, and lines of equal intensity are drawn for every 10 divisions. Clearly a small deviation in the direction of the beam would cause a serious drop in the galvanometer reading, so steps were taken to make the searchlight post as rigid as possible. The alignment of the beam was checked from time to time during the investigation, but it did not change appreciably.

Measurements with the Gold visibility meter were made on three 15 watt visibility lights at 130, 540, and 1,240 yards. Visual estimates of the visibility made during twilight were with the aid of various objects at the following distances: 175, 334, 550, 1,000, 1,100, 2,000, 2,200, 2,500, 4,000 yards, and 4½, 8, 12, and 18 miles (the last four being hills of approximately 1,000 ft.).

3. Observation Routine.

The following routine was followed when making an observation:

/(a)

(a) It was verified that the searchlight glass and photocell lens were clean, and that the photocell box was correctly aligned so that all the light from the searchlight which fell on the lens passed through the hole in the screen.

(b) The searchlight was switched on and left on for two minutes to allow it to attain a steady brightness.

(c) The shutter was opened and the galvanometer reading noted. Usually the reading fluctuated slightly and could only be read to 0.5 divisions, but occasionally (on cloudy nights, and in foggy weather) it was steadier and a more accurate reading could be made.

(d) The searchlight was switched off and the galvanometer reading (with the shutter still open) was again noted. (This was zero on dark nights). The battery voltage as measured across the searchlight switch was also immediately noted.

One or more readings were taken during twilight on most evenings when it was not raining. The galvanometer reading with the searchlight off fell to 2 - 3 divisions by sunset on fine evenings, earlier in cloudy weather, and an observation was then possible while it was still quite light enough to estimate the visibility by eye. This estimate was made immediately before making the measurement with the photocell. Some observations during morning twilight were also made. On hazy and foggy nights observations with the photocell were made at intervals during the hours of darkness, followed immediately by a measurement of the visibility with the Gold visibility-meter.

4. The Principle of the Method.

The visibility V at any time depends chiefly on the extinction coefficient μ of the air. According to Koschmieder the two quantities are related by the equation:

$$V = \frac{3.91}{\mu}$$

The visibility may thus be determined by measuring μ, and this may be done by measuring the apparent brightness of a distant light by means of a photoelectric cell and galvanometer. The apparent brightness I is given by the equation:

$$I = I_o d^{-2} e^{-\mu d}$$

where I_o is the "candle power" of the lamp and d its distance.

A series of measurements by this method has been made, and the results are compared with values of the visibility (a) measured with the Gold visibility meter, and (b) estimated visually during twilight.

5. Method of Reducing the Observations.

To facilitate reduction of the observations a graph was drawn showing the theoretical relation between the visibility and the galvanometer reading. This is shown in Fig. 3 where the abscissae are visibilities (on a logarithmic scale) and the ordinates galvanometer readings, assuming that the reading is 100 divisions for infinite visibility, i.e. for $\mu = 0$. The curve was calculated from the equation: $I = I_o d^{-2} e^{-\mu d}$.

Substituting $I_o d^{-2} = 100$ (i.e. the reading on a perfectly clear night),

d = 334 yards, and $\mu = \frac{3.91}{V}$ (Köschmieder's formula), we have $I = 100 e^{-\frac{3.91 \times 334}{V}}$

Whence $$V = \frac{568}{\log I/100} \text{ yards or } \frac{0.3226}{\log I/100} \text{ miles}$$

There are two principal factors which can affect the power of the light, (a) the battery voltage and (b) the condition of the searchlight mirror, the tarnishing of which was latterly plainly visible. A correction for the battery voltage was applied to each galvanometer reading by adding (subtracting) 3% for every 0.1 volts below (above) 12.2 volts. (The value of this correction was determined by Dr. Atkins using apparatus in his laboratory at Plymouth.) The galvanometer readings thus corrected are shown in column 5 of Tables 1 and 2.

The general trend of the galvanometer readings pointed to a progressive deterioration in the searchlight mirror used in the experimental apparatus. This complication is unfortunate. It might have been allowed for by a progressive calibration of the apparatus against visibility determined either visually or by the Gold visibility meter, but since the chief object of the trial was to assess the reliability of the photocell apparatus (or the potential reliability of an improved apparatus) a different method was adopted. The maxima in the series of galvanometer readings (corrected for battery voltage) are plotted in Fig. 4. (Table 1 gives the five observations which were used.) Partly for convenience in this application of Koschmieder's formula these maxima are assumed to represent a visibility of 39.1 km. (25 miles). It is obvious that the smoothest curve running through these maxima is practically linear, and the deterioration in the mirror was therefore assumed to be linear. Fig. 3 shows that a visibility of 25 miles gives a galvanometer reading 97% of that for infinite visibility. The dotted line of Fig.4 is 3% above the full line drawn through the plotted maxima, and therefore gives the day to day variation of the galvanometer reading for infinite visibility. This is used in the calculations of Table 2.

6. Experimental Results.

The figures in Table 2 are: (1) date, (2) time, (3) battery voltage, (4) galvanometer reading (after subtracting the reading with the searchlight off), (5) galvanometer reading corrected for battery voltage, (6) galvanometer reading for infinite visibility (from the dotted line of Fig. 4), (7) column 5 as a percentage of column 6, (8) visibility by the photocell method (obtaind from the curve of Fig. 3 at the ordinate given by the figure in column 7). For comparison there are also shown (9) the visibility estimated visually, (10) visibility measured by the Gold meter, (11) visibility measured with the Gold meter but corrected to Koschmieder's formula. (This is referred to later). The letters prefixed to the figures in column 10 indicate which of the visibility lights was used with the Gold meter on each occasion. (This was in each case the most distant one visible, but in the cases marked with an asterisk account was also taken of the fact that the next more distant light was invisible at the time.) A = 1,240 yards, B = 540 yards, C = 130 yards.

7. Discussion of Results.

A comparison between visibilities measured by different methods is shown in Fig. 5, where the ordinates are visibilities by the photocell method and abscissae are visibilities estimated visually (small circles) or measured with the Gold meter (45° crosses). The 90° crosses are referred to later. The diagonal line shows equality between ordinates and abscissae, and the distance from this line of each point represents the discrepancy in the visibility measurement.

Part of the discrepancy in the Gold meter readings is due to the different formulae used in the two cases. The formula used with the Gold meter is (Q.J. April 1939, p 144, equation (v)):

$$\beta = \frac{134}{V^{1.1}} \quad \text{or} \quad V = \left(\frac{134}{\beta}\right)^{.909}$$

where β, the "turbidity" is defined by the equation:

$$I = I_0 d^{-2} x^{\beta d}$$

(x=0.933, log x = -0.03.)

Comparing the equation: $I = I_o d^{-2} e^{-\mu d}$

we have $x^{\beta d} = e^{-\mu d}$ or $\beta \log x = -\mu \log e$

Therefore $\beta = \mu \times \frac{.4343}{.03} = 14.5 \times \mu$

Thus the formula used for the Gold meter is: $V = \left(\frac{134}{14.5\mu}\right)^{.909} = \left(\frac{9.24}{\mu}\right)^{.909}$

compared with Koschmieder's formula: $V = \frac{3.91}{\mu}$

The two formulae are shown graphically in Fig. 6, where the abscissae are visibilities on a logarithmic scale, and the ordinates are the absorption coefficient μ. It is seen that for visibilities below about three miles the Gold formula gives higher values of the visibility than Koschmieder, by one or two hundred yards in most cases. A correction was therefore applied to all the Gold visibilities to convert them to the values which would be given by Koschmieder's formula. These converted visibilities are shown in column (11) of Table 2, and by 900 crosses in Fig. 5.

8. Conclusions.

Up to about 2,000 yards the visibility by photo-cell apparatus tends to be higher by some 10-20% than the visibility estimated visually. The photo-cell method gives values about 30% lower than the Gold meter method up to about 2,000 yards, but the difference is somewhat reduced if the Gold meter values are corrected as indicated above. For visibilities between about 2,000 yards and three miles the photo-cell, visual and Gold meter methods give on the average about the same values with no undue scatter. On the whole the differences as between the different methods are no greater than the discrepancies found in estimating visibility in daylight by visual methods.

For visibility above about three miles there is an increasing discrepancy, the photocell method giving values lower than the visual estimates. One reason for this is that distant hills are observed partly through the clearer layers of air at some height above the ground. But the discrepancy may be to some extent inherent in the method of reducing the observations. For if the exceptionally high galvanometer readings used in plotting Fig. 4 were due to some instrumental reason rather than to exceptionally clear air the effect would be to make the calculated visibilities too low, the error increasing rapidly with increasing visibility above about 5 miles.

If this apparatus is to become a standard piece of M.O. equipment the following practical suggestions are made:

(a) The lamp should be run from the mains via a voltage stabilizer.
(b) The mirror should be of a metal having a stable reflection coefficient.
(c) If possible the beam should be broadened, perhaps by a very thin diffusing screen in front of the searchlight, so that a small shift in the axis of the beam would not make such a large difference to the galvanometer reading. Nevertheless the searchlight should be mounted as rigidly as possible.
(d) The searchlight should be protected from rain and dew by a cowl, and should preferably have a small heater (switched on with the lamp) to ensure that the glass is dry. A desiccator inside the searchlight would also be desirable, to prevent condensation on the mirror.
(e) It is equally important that the photocell lens should be kept dry, but that would usually be more accessible from the office, and a short lens hood should suffice.

/(f)

(f) The Tinsley galvanometer is very satisfactory, and should be retained; it is easy to read in poor light, and parallax errors are impossible. It is suggested that a scale could be fitted to show the visibility directly, if a fairly constant searchlight was used. Small variations in the latter could be allowed for by varying the sensitivity of the galvanometer, e.g. by a variable high resistance in parallel with it. The procedure would be that on clear evenings the galvanometer would be adjusted by the parallel resistance to read the actual estimated visibility; then when the visibility was poor the galvanometer would show the value of the visibility directly. With the lamp at its present distance (334 yards), the scale would be as shown below:

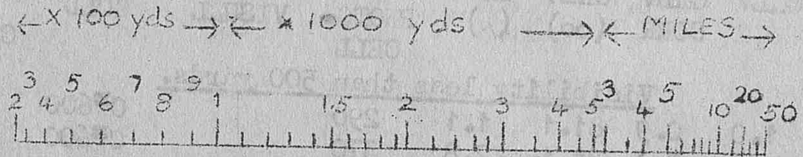


TABLE 1.

(1) DATE 1944	(2) TIME GMT	(3) VOLTS	(4) GALV.	(5) GALV. CORR	(9) VISIBIDITY VISUAL.
10/8	2005	12.15	86	87.3	15 miles.
23/8	1945	12.0	79	83.8	20
23/9	1900	12.45	85	78.6	15
14/10	1720	12.3	77	74.7	25
23/10	1705	12.2	73.5	73.5	30

TABLE 2.

(1) DATE 1944	(2) TIME GMT	(3) VOLTS	(4) GALV	(5) GALV CORR	(6) GALV (∞)	(7) GALV (%)	(8) VISIBILITY (YARDS) PHOTO-CELL	(9) VISUAL	(10) GOLD	(11) GOLD CORRECTED
Visibility less than 500 yards.										
22/9	0150	12.5	1.0	0.9	81.1	1.1	290		C*600	480
22/9	0235	12.5	2.5	2.3	81.1	2.8	370		C*600	480
22/9	0649	12.5	3.5	3.2	81.1	3.9	400	350		
23/10	0245	12.2	0.0	0.0	74.8	0.0	< 200		C 240	180
23/10	0352	12.2	1.0	1.0	74.8	1.3	310		C 400	310
23/10	0409	12.2	2.5	2.5	74.8	3.3	390		B 600	480
26/10	0729	12.15	0.6	0.6	74.2	0.8	270	200		
26/10	0743	12.15	0.6	0.6	74.2	0.8	270	250		
26/10	0755	12.15	0.7	0.7	74.2	0.9	280	250		
Visibility 500 - 1,000 yards.										
22/9	0645	12.5	6.5	5.9	81.1	7.3	500	550		
22/9	0706	12.5	9.0	8.2	81.1	10.1	570	600		
22/9	0718	12.5	23.5	21.4	81.1	25.4	980	750		
22/10	2330	12.2	18.2	18.2	75.0	24.3	930		B 1,350	1,200
23/10	0415	12.2	9.7	9.7	74.8	13.0	640		B 900	750
23/10	0421	12.2	12.2	12.2	74.8	16.3	720		B 1,050	900
23/10	0427	12.2	18.0	18.0	74.8	24.1	920		B 1,200	1,040
26/10	0910	12.2	6.2	6.2	74.2	8.4	530	400		
29/10	0741	12.1	16.5	17.0	73.6	23.1	900	900		
2/11	0729	12.1	9.5	9.8	72.8	13.5	650	500		
2/11	0735	12.1	10.5	10.8	72.8	14.8	690	600		
2/11	0811	12.1	6.5	6.7	72.8	9.2	550	550		
2/11	0824	12.15	7.5	7.6	72.8	10.4	580	600		
2/11	0837	12.15	11.8	12.0	72.8	16.5	720	650		
2/11	0858	12.15	14.0	14.2	72.8	19.5	800	800		
2/11	0904	12.1	15.7	17.2	72.8	23.6	900	900		
2/11	0923	12.1	16.0	16.5	72.8	22.7	880	900		
Visibility 1,000 - 2,000 yards.										
22/9	0724	12.5	29	26.4	81.1	32.5	1,170	900		
22/9	0730	12.5	30.5	27.8	81.1	34.3	1,230	1,000		
8/10	1741	12.35	42	41.1	77.8	51.6	1,970	1,500		
21/10	0738	12.2	37.5	37.5	75.2	49.9	1,880	2,500		
21/10	0815	12.2	38.5	38.5	75.2	51.2	1,930	3,000		
22/10	0726	12.2	31.5	31.5	75.0	42.0	1,510	2,000		
22/10	0757	12.25	30.5	30.0	75.0	40.0	1,430	1,700		
22/10	2232	12.2	33.5	33.5	75.0	44.6	1,630		A 2,000	1,850
22/10	2255	12.2	26	26.0	75.0	34.6	1,240		A 1,500	1,300
23/10	0006	12.2	26	26.0	74.8	34.8	1,250		B*1,400	1,200

TABLE 2 (continued)

(1) DATE 1944	(2) TIME GMT	(3) VOLTS	(4) GALV	(5) GALV CORR	(6) GALV (∞)	(7) GALV (%)	(8)	(9)	(10)	(11)
							VISIBILITY (yards)			
							PHOTO-CELL	VISUAL	GOLD	GOLD CORRECTED
				Visibility 1,000 - 2,000 yards (continued)						
23/10	0054	12.2	21	21.0	74.8	28.1	1,030		A 1,380	1,200
23/10	0146	12.2	23.5	23.5	74.8	31.4	1,120		A 1,400	1,200
29/10	0810	12.1	28	28.8	73.6	39.1	1,400	1,200		
29/10	0816	12.1	31	31.9	73.6	43.3	1,580	1,700		
29/10	0828	12.1	35.5	36.5	73.6	49.6	1,870	2,000		
29/10	0836	12.1	33	34.0	73.6	46.2	1,710	2,200		
29/10	0850	12.1	32.5	33.5	73.6	45.5	1,680	1,300		
29/10	0857	12.1	32.5	33.5	73.6	45.5	1,680	2,000		
29/10	0915	12.1	31	31.9	73.6	43.4	1,580	2,300		
29/10	0926	12.1	31.5	32.4	73.6	44.0	1,600	2,500		
1/11	0732	12.1	25	25.7	73.0	35.2	1,260	1,000		
1/11	0814	12.15	24.5	24.9	73.0	34.1	1,220	1,000		
1/11	0851	12.15	22.5	22.8	73.0	31.2	1,120	1,000		
2/11	1902	12.1	36.5	37.6	72.8	51.6	1,980		A 2,350	2,200
3/11	0150	12.1	31	31.9	72.6	44.0	1,600		A 2,300	2,150
				Visibility 2,000 - 3,000 yards.						
14/9	1900	11.9	48	52.3	82.8	63.2	2,800	2,500		
21/9	2353	12.5	52	47.3	81.4	58.1	2,400		A 2,800	2,600
22/9	0053	12.5	50.5	45.9	81.2	56.5	2,300		A 2,800	2,600
6/10	0058	12.4	52.5	49.4	78.3	63.1	2,800		A 3,500	3,300
8/10	0810	12.35	51	48.7	77.9	62.5	2,750	3,300		
8/10	0858	12.35	50	47.7	77.9	61.2	2,650	4,000		
8/10	1653	12.35	44.5	42.5	77.8	54.6	2,200	2,500		
8/10	1715	12.35	44.5	42.5	77.8	54.6	2,200	2,300		
8/10	1733	12.35	43	41.1	77.8	52.8	2,050	1,900		
8/10	1752	12.35	49	46.8	77.8	60.1	2,600	2,800		
9/10	0635	12.35	50.5	48.2	77.7	62.0	2,700	3,700		
9/10	0712	12.35	40	47.7	77.7	61.4	2,650	3,300		
9/10	0727	12.35	49	46.8	77.6	60.3	2,600	3,500		
10/10	0728	12.3	47	45.6	77.5	58.9	2,500	2,800		
10/10	0740	12.3	47.5	46.1	77.5	59.5	2,500	2,300		
10/10	0805	12.3	45.5	44.1	77.5	56.9	2,300	2,500		
22/10	1913	12.25	43.5	42.8	75.0	57.1	2,300		A 2,500	2,300
22/10	2051	12.25	44.5	43.8	75.0	58.4	2,400		A 2,300	2,150
23/10	0502	12.2	43	43.0	74.8	57.5	2,350		A 2,500	2,300
23/10	0647	12.25	45	44.3	74.8	59.2	2,500	2,000		
23/10	0702	12.25	45	44.3	74.8	59.2	2,500	2,500		
31/10	1700	12.15	40.5	41.1	73.2	56.1	2,250	2,200		
2/11	1806	12.15	44.5	45.2	72.8	62.1	2,750		A 2,650	2,450
2/11	2034	12.1	44	45.3	72.8	62.2	2,750		A 3,500	3,300
2/11	2342	12.1	41	42.2	72.8	58.0	2,400		A 3,400	3,200
3/11	0721	12.15	43	43.6	72.6	60.0	2,550	3,300		
				Visibility 3,000 - 4,000 yards.						
14/9	1830	11.85	53.5	59.1	82.8	71.4	3,900	3,000		
14/9	1848	11.9	50	54.5	82.8	65.9	3,150	2,800		
21/9	2153	12.5	58.5	53.2	81.4	65.3	3,050		A 3,100	2,900
6/10	0250	12.4	58	54.5	78.3	69.3	3,600		A 5,700	5,700
8/10	0740	12.35	54	51.6	77.9	66.3	3,200	3,300		

TABLE 2 (continued)

(1) DATE 1944	(2) TIME GMT	(3) VOLTS	(4) GALV	(5) GALV CORR	(6) GALV (∞)	(7) GALV (%)	(8)	(9)	(10)	(11)
								VISIBILITY		
							PHOTO-CELL	VISUAL	GOLD	GOLD CORRECTED
8/10	0828	12.35	56.5	54.0	77.9	69.4	3,600	3,600		
8/10	0840	12.35	52.5	50.1	77.9	64.3	3,000	4,000		
9/10	0450	12.35	57	54.4	77.7	70.0	3,700		A 5 miles	5¼miles
9/10	0605	12.35	56.5	54.0	77.7	69.5	3,600	3,700		
22/10	1956	12.25	50	49.2	75.0	65.6	3,050		A 2,600	2,400
24/10	1718	12.2	52	52.0	74.6	69.7	3,600	2,800		
				Visibility 4,000 - 5,000 yards.						
8/10	1850	12.4	63	59.2	77.9	76.0	4,750		A 5 miles	5¼miles
9/10	0010	12.35	62	59.2	77.6	76.4	4,850		A 4	4
22/10	1650	12.25	56.5	55.7	75.0	74.3	4,500	3,500		
22/10	1706	12.25	57	56.1	75.0	74.8	4,500	3,500		
24/10	1706	12.15	53	53.8	74.6	72.1	4,000	2,500		
26/10	1652	12.15	56	56.8	74.2	76.5	4,900	3,500		
1/11	1652	12.1	51	52.5	73.0	71.9	4,000	6,000		
2/11	1700	12.1	52	53.5	72.8	73.5	4,250	3,500		
				Visibility 3 - 6 miles.						
19/8	1950	12.05	70	73.2	88.0	83.2	4.0	5		
19/8	2000	12.05	71.5	74.7	88.0	84.9	4.5	5		
6/9	1925	11.95	69	74.1	84.5	87.6	5.5	12		
10/9	1910	11.95	67.5	72.5	83.5	86.8	5.3	12		
12/9	1855	11.9	65.5	71.4	83.1	84.8	4.5	12		
13/9	2053	11.9	59.5	64.8	83.0	78.1	3.0	4		
21/9	1835	12.45	75.5	69.8	81.4	85.8	4.9	9		
21/9	1915	12.5	73.5	66.9	81.4	82.1	3.7	7		
22/9	1745	12.5	69.5	63.2	81.1	78.0	3.0	5		
22/9	1806	12.5	71.5	65.1	81.1	80.3	3.3	5		
22/9	1825	12.5	70.5	64.1	81.1	79.0	3.1	3.5		
27/9	1818	12.45	76	70.3	80.1	87.8	5.6	20		
29/9	1734	12.45	76	70.3	79.7	88.1	5.9	15		
4/10	1851	12.35	68.5	65.4	78.7	83.1	4.0	5		
5/10	0600	12.35	72	68.8	78.5	87.6	5.5	10		
5/10	0615	12.35	71	67.8	78.5	86.4	5.1	9		
5/10	0633	12.35	69.5	66.7	78.5	84.6	4.4	9		
5/10	1740	12.35	70.5	67.3	78.5	85.7	4.8	11		
5/10	1818	12.4	72.5	68.1	78.5	86.8	5.3	10		
6/10	0605	12.4	64.5	60.6	78.2	77.5	3.0	3.5		
6/10	0615	12.35	66	63.0	78.2	80.5	3.3	5		
6/10	0655	12.35	64	61.1	78.2	78.1	3.0	6		
7/10	1555	12.35	71.5	68.2	78.1	87.4	5.5	13		
9/10	1703	12.35	67.5	64.5	77.6	83.1	4.0	5		
9/10	1726	12.35	66.5	63.5	77.6	81.9	3.7	3.5		
13/10	1745	12.25	68.2	67.2	76.8	87.5	5.5	14		
21/10	1623	12.25	63.5	63.5	75.2	84.4	4.4	2.8		
25/10	1642	12.2	64	64.0	74.4	86.0	5.0	3		
25/10	1655	12.2	62	62.0	74.4	83.3	4.0	3.5		
				Visibility 6 - 10 miles.						
13/8	2004	12.15	81	82.3	89.3	92.2	9.3	20		
18/8	1955	12.05	77.8	81.3	88.2	92.2	9.3	12		
8/9	1907	11.95	71.5	76.8	84.0	91.4	8.3	30		
19/9	1845	12.45	78.5	72.5	81.7	88.7	6.3	11		
25/9	1816	12.45	77	71.1	80.5	88.4	6.2	12		

(1) DATE 1944	(2) TIME GMT	(3) VOLTS	(4) GALV	(5) GALV CORR	(6) GALV (∞)	(7) GALV (%)	(8) PHOTO-CELL	(9) VISIBILITY VISUAL	(10) GOLD	(11) GOLD CORRECTED
28/9	1802	12.4	76.3	71.7	79.9	89.8	6.9	15		
29/9	1759	12.45	76.5	70.8	79.7	88.8	6.3	13		
1/10	1830	12.4	76.5	71.9	79.3	90.6	7.5	20		
6/10	1755	12.4	74.5	70.0	78.3	89.4	6.6	7.5		
7/10	1725	12.35	74	70.6	78.1	90.5	7.5	11		
10/10	1645	12.3	71	68.9	77.5	89.0	6.4	6.5		
10/10	1704	12.3	71	68.9	77.5	89.0	6.4	5.5		
12/10	1745	12.3	73	70.8	77.0	92.0	9.0	25		
13/10	1730	12.3	72.5	70.3	76.8	91.5	8.5	16		
3/11	1645	12.1	65	67.0	72.6	92.3	9.4	18		

Visibility over 10 miles.

9/8	2005	12.15	84	85.3	90.1	94.6	15	20
12/8	2010	12.15	83	84.3	89.5	94.2	13	30
17/8	2015	12.05	79.5	83.1	88.4	94.0	12	15
21/8	1948	12.0	78.5	83.3	87.6	95.0	15	15
6/9	1905	11.9	71.5	78.0	84.5	92.5	10	12
15/9	1830	11.85	70.5	78.0	82.6	94.5	13	18
15/9	1845	11.85	70.	77.4	82.6	93.7	11	15
15/9	1855	11.9	70.2	77.6	82.6	94.0	12	16
23/9	1914	12.45	82.8	76.5	80.9	94.6	14	15
24/9	1843	12.45	82.5	76.3	80.7	94.5	13	15
12/10	1737	12.3	74	71.8	77.0	93.2	11	20
14/10	1734	12.3	73.7	71.5	76.6	93.4	11	22
14/10	1742	12.3	73.5	71.3	76.6	93.0	10	20
19/10	1710	12.25	73.2	72.1	75.5	95.5	17	15
27/10	1655	12.15	71	72.2	74.0	97.6	32	30
3/11	1658	12.1	65.5	67.5	72.6	93.0	10	20
5/11	1714	12.1	67	69.0	72.1	95.7	17	18
6/11	1655	12.1	67	69.0	71.9	96.0	18	15
7/11	0731	12.05	65.5	68.5	71.7	95.4	16	18

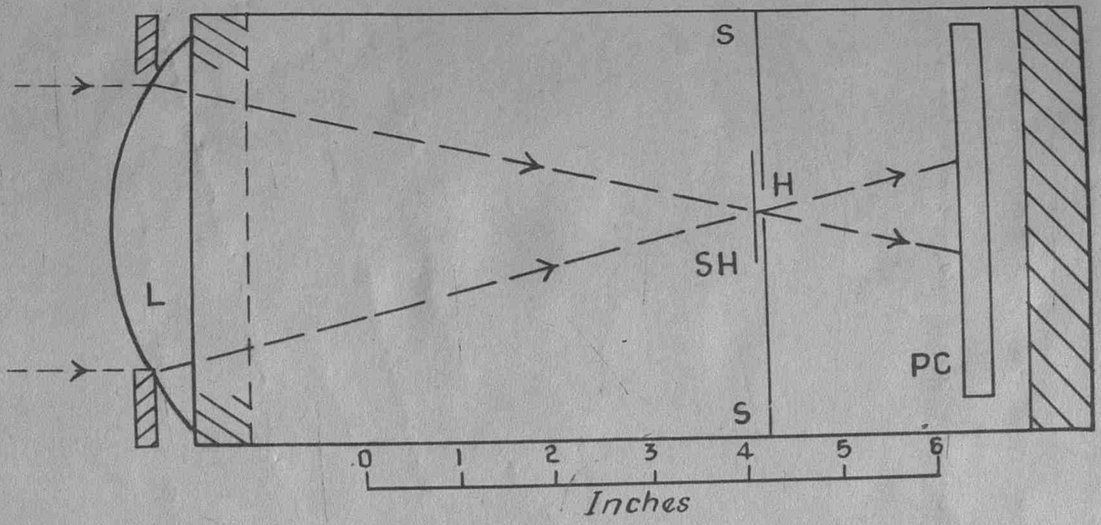

S
H
SH
L
PC
S
0
1
2
3
4
5
6
Inches

FIG. 1.

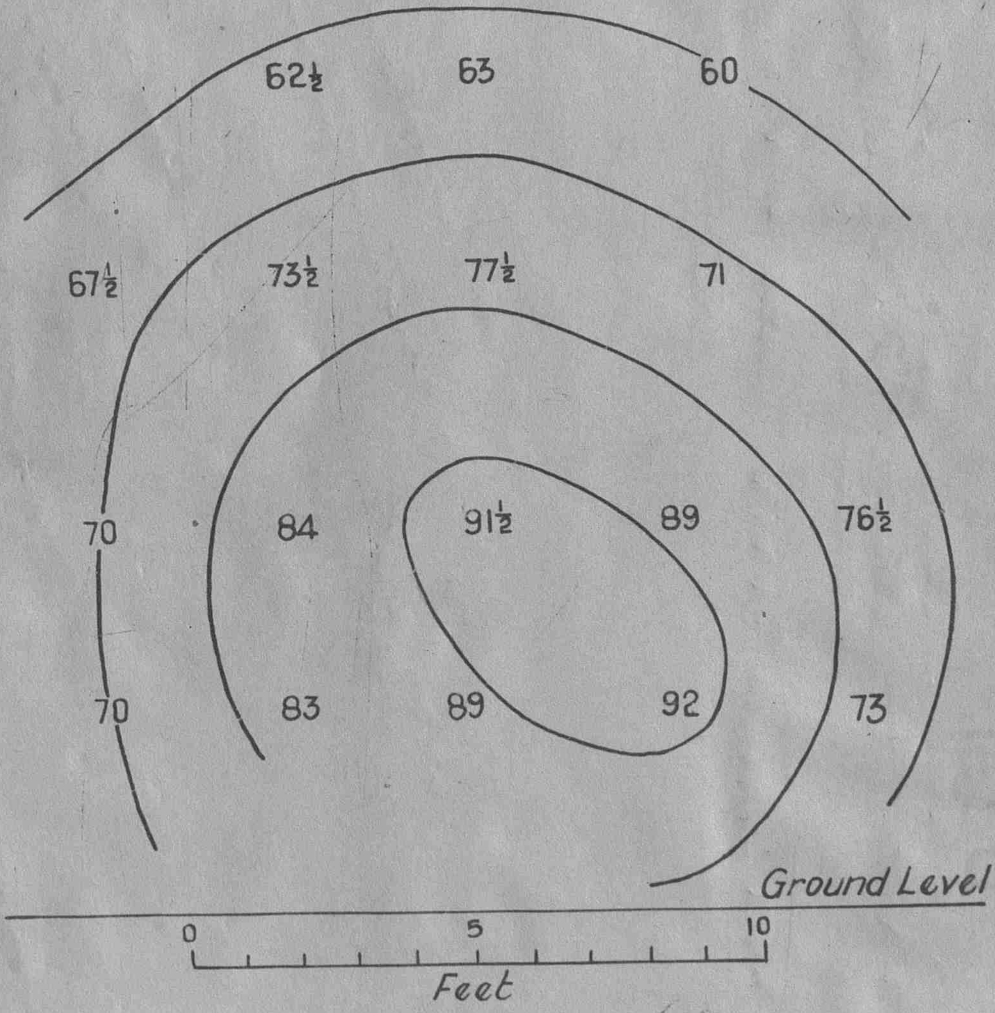
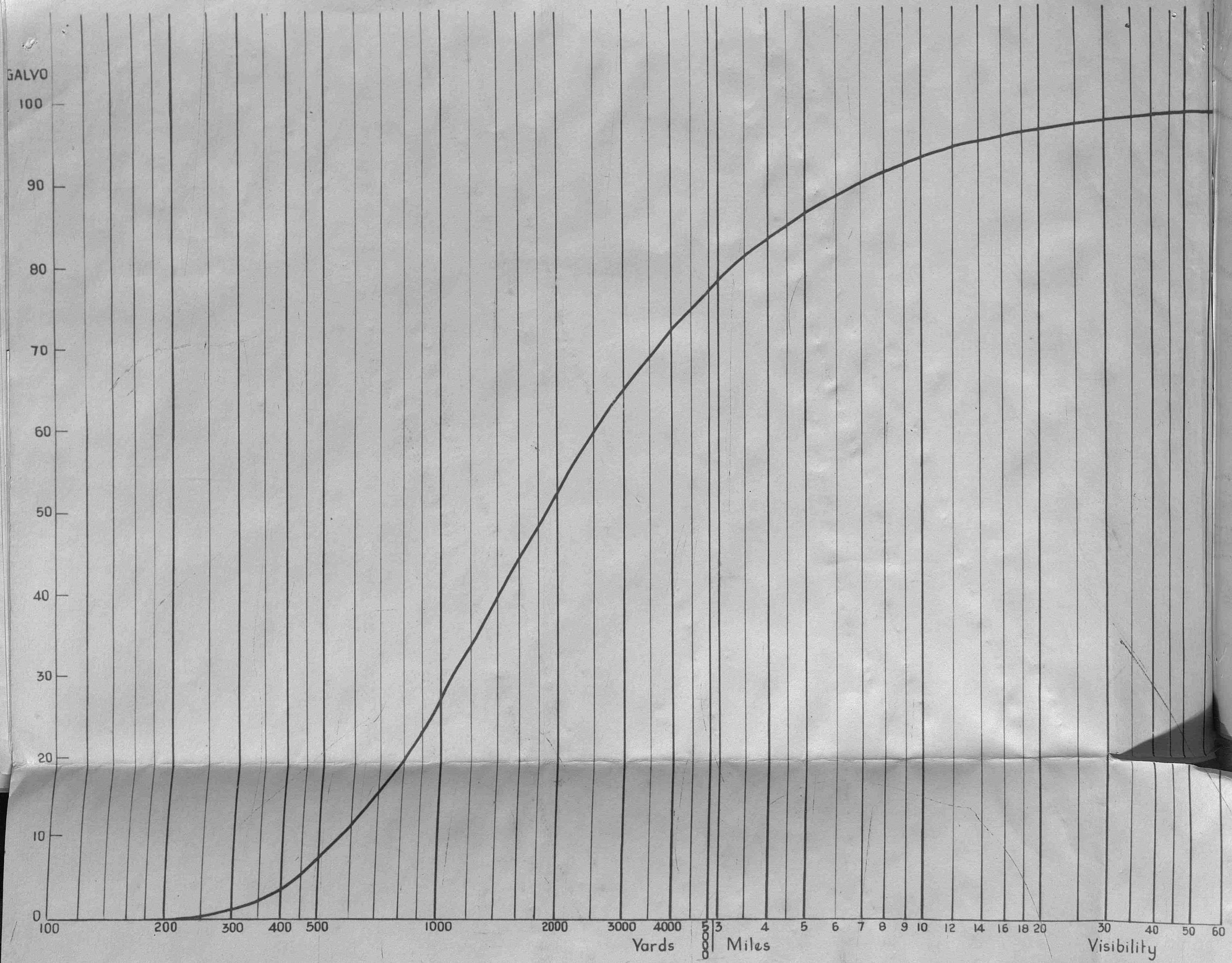

62½
63
60
67½
73½
77½
71
70
84
91½
89
76½
70
83
89
92
73
Ground Level
0
5
10
Feet

FIG. 2.

FIG. 3.


GALVO
100
90
80
70
60
50
40
30
20
10
0
100
200
300
400
500
1000
2000
3000
4000
5000
3
4
5
6
7
8
9
10
12
14
16
18
20
30
40
50
60
Yards
Miles
Visibility

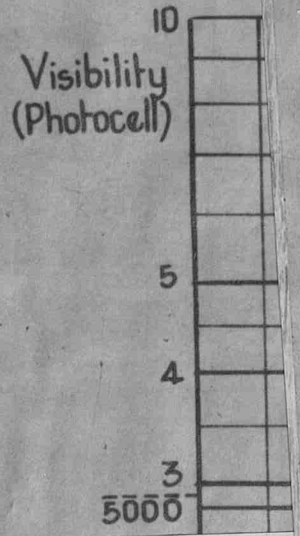

Visibility
(Photocell)
10
5
4
3
5000

FIG. 4.

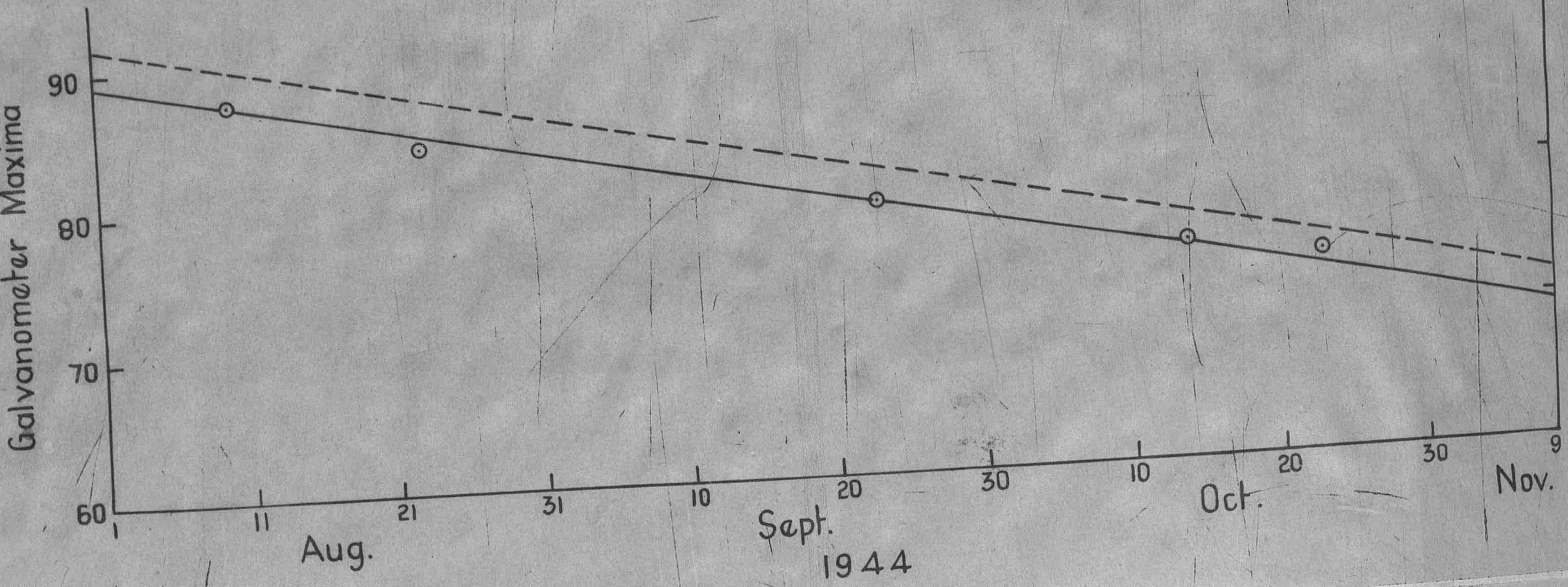
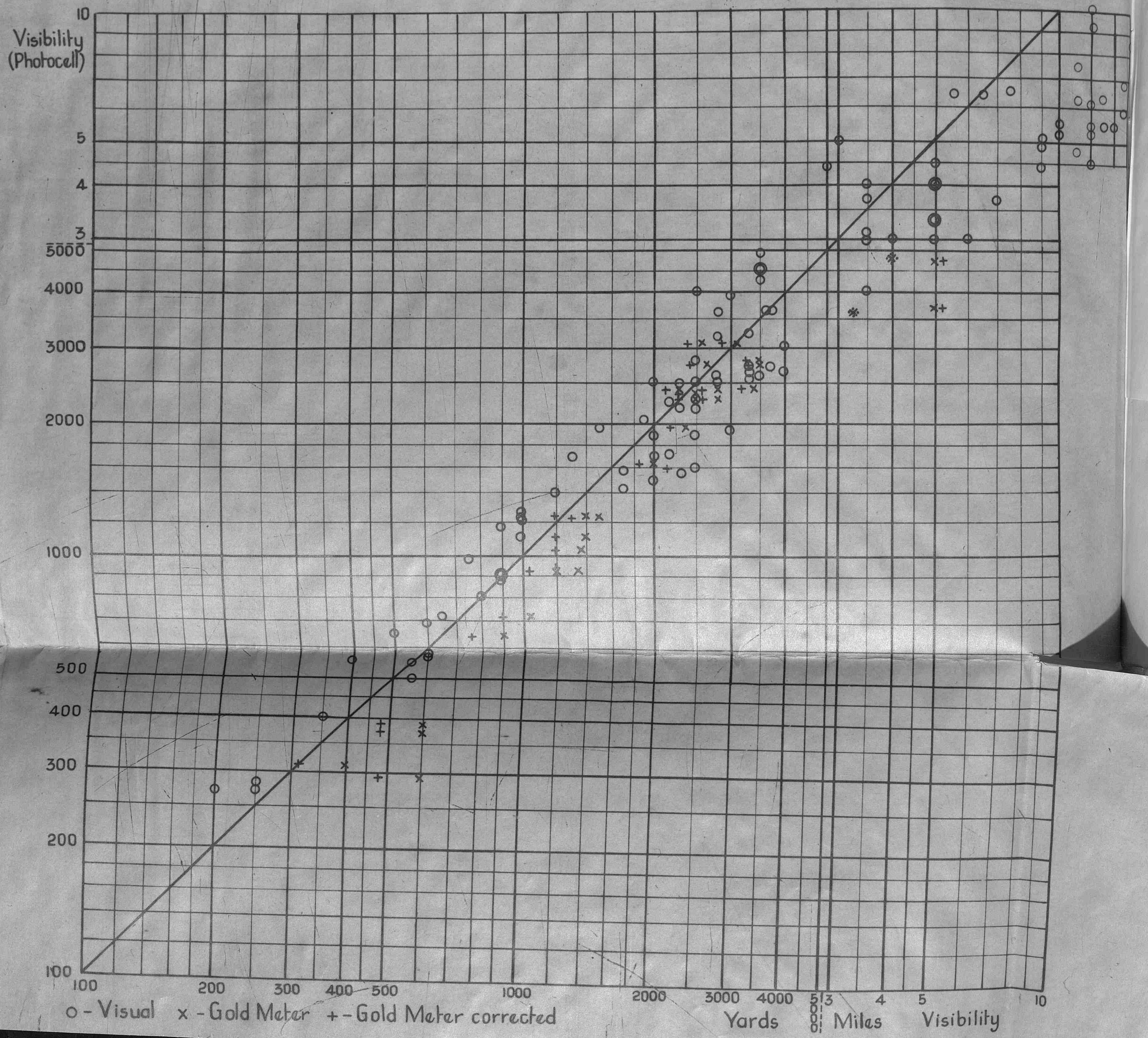

Galvanometer Maxima
90
80
70
60
1
11
21
31
10
20
30
10
20
30
9
Aug.
Sept.
Oct.
Nov.
1944

FIG. 5.


Visibility (Photocell)
10
5
4
3
5000
4000
3000
2000
1000
500
400
300
200
100
100
200
300
400
500
1000
2000
3000
4000
5000
3
4
5
10
Yards
Miles
Visibility
o - Visual
x - Gold Meter
+ - Gold Meter corrected

RESTRICTED

MRP 24

J.M.R.P. No.17.

METEOROLOGICAL INVESTIGATIONS AT RYE

Part I. Instrumental Layout for Recording Gradients of Temperature and Relative Humidity.

1. The Site. The geographical coordinates of the Rye Station are: Latitude 50° 58' 10" N, Longitude 0° 48' 50" E. The site is on level grassland about 12ft. above M.S.L., the nearest hills being over 4 miles away to the west and northwest.

A 360ft. radio lattice-work tower, a photograph of which is shown in Fig. 1, is fitted on the W.S.W. side with small Stevenson screens at heights of 50ft., 155ft. and 350ft. The effect of the tower on the meteorological elements is considered to be negligible. The screens are mounted on angle-iron rails extending $5\frac{1}{2}$ft. out from the side of the tower. For convenience in servicing, the screens are fitted with wheels and can be pulled into or away from the tower as required. The 50ft. screen is illustrated in Fig. 2.

Another Stevenson screen is erected on a concrete bed at a height of 4ft. about 10 yards from the base of the tower, on the south side.

2. Temperature Recording Apparatus. The following quantities are recorded:-

Air temperature at 4ft.
Difference in temperature between 50ft. and 4ft.
" " " " " " 155 "
" " " " " " 350 "

(a) Element. The thermometer elements consist of several turns of nickel wire encased in a brass sleeve (Fig.3). The relation between the temperature and the resistance of the element is shown in the following table:-

Temp. °F.	Resist. (ohms)
-30	84.4
-20	86.8
-10	89.3
0	91.8
10	94.3
20	96.9
30	99.5
32 (Ice point)	100
40	102.1
50	104.7
60	107.4
70	110.1
80	112.8
90	115.6
100	118.4

The elements are mounted in small Stevenson screens, as shown in Figs. 2 and 3. Under these conditions the lag coefficient, defined as the time required for the difference in temperature between the element and the surrounding air (outside the screen) to be reduced to $\frac{1}{e}$ of its original value, varies from roughly $3\frac{1}{2}$ minutes to 6 minutes for wind speeds outside the screen of 25 m.p.h. and 5 m.p.h. respectively. This means that these elements are only suitable for measuring slow period changes in temperature. The above results were obtained by heating the thermometer element to a temperature T, about 20°F. above the air temperature T^1 and measuring the time it took to reach a temperature $T^1 + \frac{T - T^1}{e}$ when the source of heat was suddenly removed.

/ (b) Circuit

(b) Circuit. The electrical circuits are shown diagramatically in Figs. 4 and 5. They are, fundamentally, out-of-balance Wheatstone networks. In the differential temperature circuits, two arms of the bridge are resistance thermometers.

Compensation leads are provided in the case of the differential temperatures, each lead being taken to the 350ft. level before being connected to the recorder. No compensation is necessary for the 4ft. temperature recorder, as the leads are relatively short.

All the external leads are Pyrotenax cable, which consists of a copper core, surrounded by a plastic insulator, and enclosed in a copper tube. This cable provides good external earthing and screening combined with good internal insulation.

(c) Recorders. The recorders, illustrated in Figs. 6 and 7, are of the Negretti & Zambra electrical thermometer type. The position of the galvanometer needle is recorded on a chart by the action of a chopper, which presses the needle at half-minute intervals against an ink ribbon in contact with the chart. The recorders are electrically operated from the mains.

The galvanometer is a moving coil instrument, spring controlled and double pivoted, with spring jewelled bearings. It is provided with zero adjustment. The movement is totally enclosed in a separate box, mounted on shock absorbers inside the main case.

The switching mechanism is fitted with cam and pawl operated mercury tubeswitches, which give a fine positive action and minimise the risk of variable contact resistance.

The chart is driven by twin driving wheels acting through the perforated edges of the chart. The ink ribbon (in the case of the differential recorder a 3-coloured ribbon is used) is automatically wound from one end to the other. The time scale is ½" per hour and the charts run for 50 days.

The 4ft. actual temperature recorder is a single-point instrument with a range of 0-100°F. The scale value is approximately 20°F. per inch, and it is possible to estimate the temperature to 1/4°F.

A three-point recorder is used for the differential temperatures, the element at 50ft. forming one arm of the Wheatstone bridge and the elements at 4ft., 155ft. and 350ft. each being connected in turn to the adjacent arms. The differences between the temperature at 50ft. and the other three heights are thus recorded as three separate traces on one chart, which are readily distinguishable by their different colours. The range is ±10°F. with a scale value of 4°F. per inch. Temperature differences can be measured easily to 1/10°F. The maximum range has been exceeded occasionally, and arrangements are in hand for the range to be doubled when required by operating a switch.

(d) Recorder Room. The recorders are mounted as shown in Fig. 7 on a brick wall in an inner room of a service hut. The temperature of the room is thermostatically controlled at 60-62°F, and the air is kept fresh by an extractor fan.

In order to keep a check on the voltage variations and failures of the mains supply, a recording voltmeter by Evershed and Vignoles has been installed.

3. Humidity Recording Apparatus. The relative humidities at 4ft., 50ft., 155ft. and 350ft. are recorded by one set of instruments, and an independent additional recording of the humidity at 4ft. is made with another instrument.

The apparatus was developed by Messrs. Negretti & Zambra in collaboration with Prof. H. Spencer-Gregory of the Imperial College of Science and Technology, South Kensington. The Rye installation is the first application of a recording Gregory humidiometer for meteorological purposes.

/ (a). Element

(a) Element. The principle of the Gregory humidiometer element is the variation with relative humidity of the electrical resistance of a piece of fabric impregnated with a hygroscopic salt, such as lithium or calcium chloride. The device is patented (British Patent Specification No.533,381).

Fig. 8 is a photograph of one of the elements in use at Rye. The impregnated fabric strip, made of finely woven Egyptian cotton, is stretched around a framework of 24 silver rods mounted in a cylindrical Perspex holder. The rods are wired alternately in parallel, so that in effect twelve separate lengths of the fabric are electrically connected in parallel. The fabric is kept under tension by stainless steel springs to ensure good contact with the silver rods. The overall dimensions of the element are 2" diameter and 2¼" long.

The element is prepared by soaking the fabric first in benzole, then in boiling distilled water and finally in a hot 4-6% solution of lithium chloride. After this treatment it is allowed to dry before being mounted.

The element is calibrated by exposure to the atmosphere above aqueous solutions of sulphuric acid of known strength. The relation between the resistance of an element and the relative humidity is shown in the following table:-

Relative Humidity %	30	45	50	60	70	80	90	95	100
Resistance (ohms)	113,500	7,621	3,784	1,125	402	117	92	68	52

The resistance of the element is a function of temperature as well as humidity, and in order to allow for this a correction to the recorded value of the relative humidity is applied with the aid of the curves shown in Fig.9. A change of 10°C. in temperature is roughly equivalent to a change of 5% in the relative humidity at 30% or a change of 3% at 90%.

Unlike most hygrometers, the response of the element to changes of relative humidity follows an approximately exponential law given by the equation

$$\frac{dH}{dt} = -\frac{1}{\lambda}\,(H - H^1)$$

where H is the instantaneous reading of the humidiometer at time t, H^1 is the actual relative humidity of the medium and λ is a constant, known as the lag coefficient, with the dimensions of time. For a ventilation speed of 5 m.p.h., and over a humidity range of 85% to 55%, the lag coefficient of a Gregory element was found to be 90 seconds.

The above results were obtained with an element fully exposed to the airflow in a wind tunnel. The elements at Rye are mounted in metal cases, about 3½" cube, with shielded openings in the side, see Fig.3, which are in turn mounted in Stevenson screens. It has not been possible to measure the lag under these circumstances, but it must be appreciably greater than the wind tunnel results. The provision of artificial aspiration is being considered with a view to making the response more rapid.

(b) Circuit. The electrical circuits are shown diagrammatically in Figs. 10 and 11. An alternating current from a voltage stabilising unit is passed through the element, and after rectification the changes in current are recorded by a galvanometer recorder.

As the changes in resistance are so large, no compensation for the resistance variations of the leads is required.

/ (c) Recorders

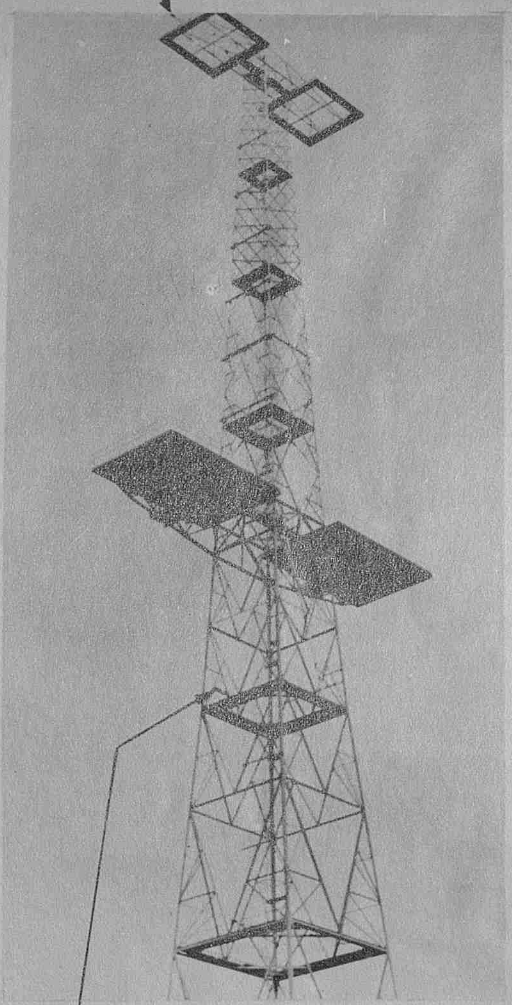
(c) Recorders. The recorders are of the same type as those used for temperature, as described in section 2(c). The humidities at the four different heights are recorded as different coloured traces on a single chart by means of a four point recorder, while the second 4ft. element is connected to a single point recorder. Full deflection of the galvanometer needle, corresponding to 100% relative humidity, is produced by a current of roughly 0.4 milliamps.

As the resistance of the Gregory element is not a linear function of the relative humidity, the scale of the record is not uniform. The scale value corresponding to a 1% change of humidity increases from about .03 inch at 40% to .13 inch at 65% and decreases again to .03 inch at 95%. Over the open part of the scale the humidity can be measured to 0.1% but the accuracy is of the order of ±1%. The lower limit of the present recorders is 30%, but the provision of an outfit to record still lower is being considered.

The recorders are mounted alongside the temperature recorders, as shown in Fig.7 and described in section 2(d).

Instruments Branch,
Meteorological Office (M.O.4),
Air Ministry.
May, 1944.

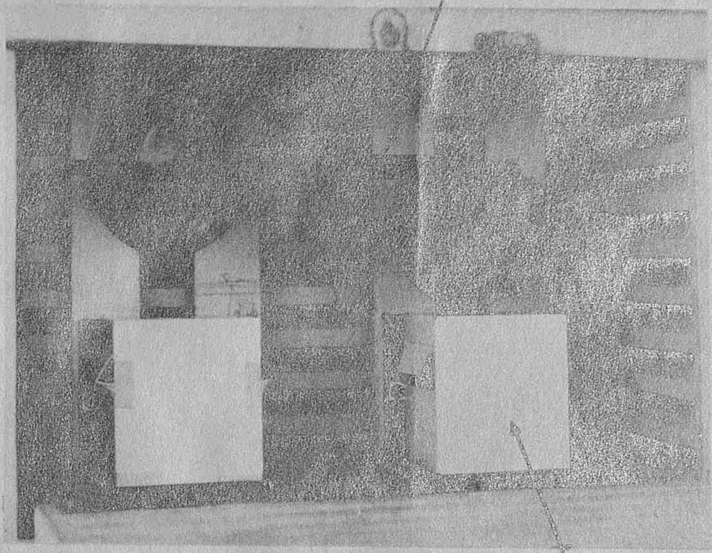
350 ft Screen.



155 ft Screen.

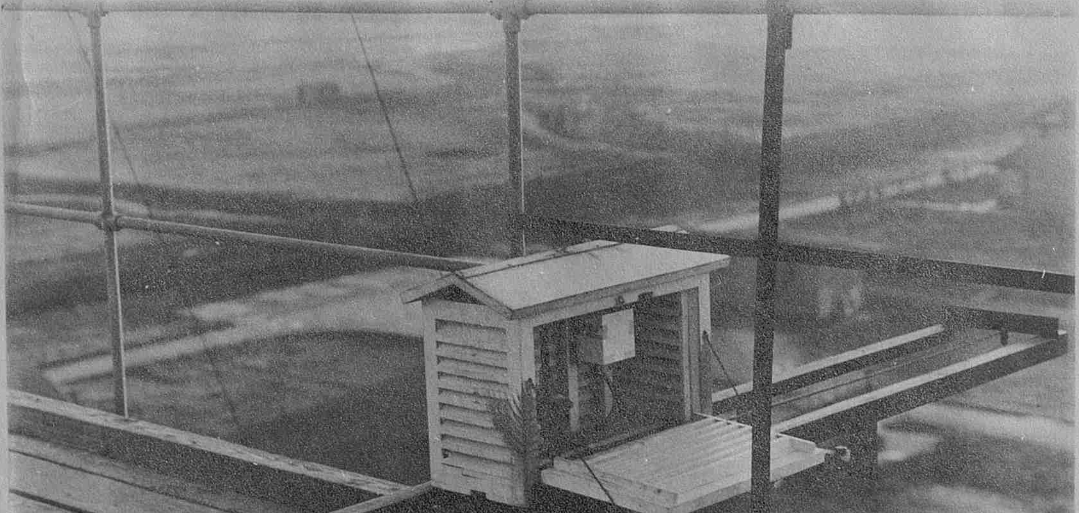
FIG. 1. RYE TOWER.

Resistance Thermometer.



Housing of Humidiometer Element.

FIG. 3. INTERIOR OF 4 FT STEVENSON SCREEN.



Pyrotenax Cable.

FIG. 2. 50 FT SCREEN.

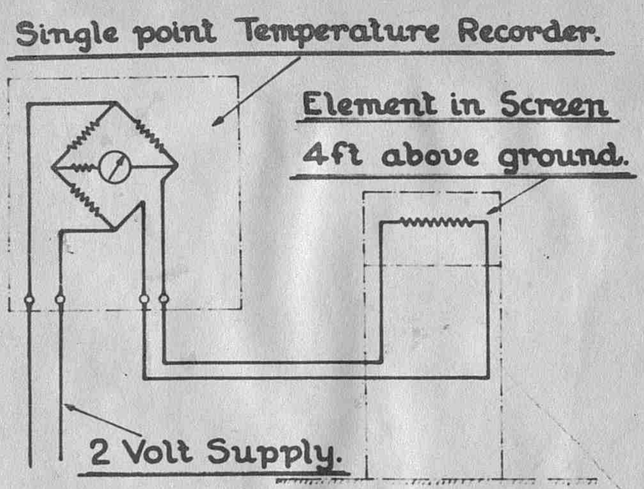

Single point Temperature Recorder.
Element in Screen
4ft above ground.
2 Volt Supply.

FIG. 4. MEASUREMENT OF ACTUAL TEMPERATURE AT 4FT.

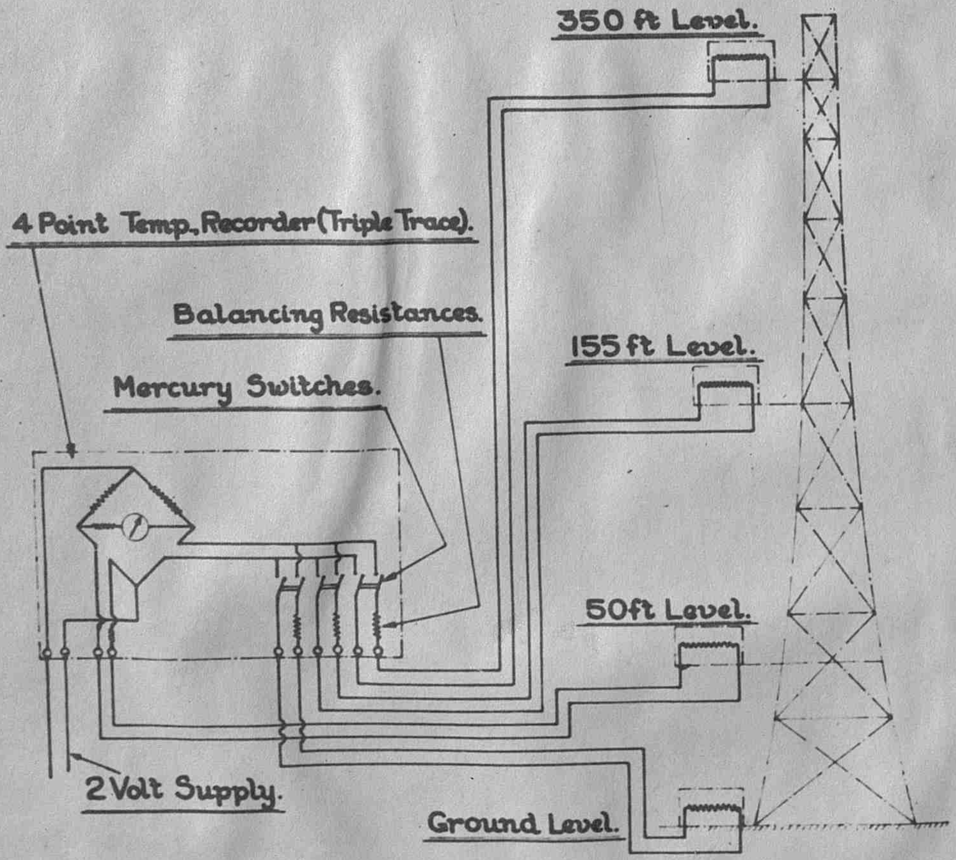

350 ft Level.
4 Point Temp. Recorder (Triple Trace).
Balancing Resistances.
155 ft Level.
Mercury Switches.
50ft Level.
2 Volt Supply.
Ground Level.

FIG. 5. MEASUREMENT OF DIFFERENTIAL TEMPERATURES.

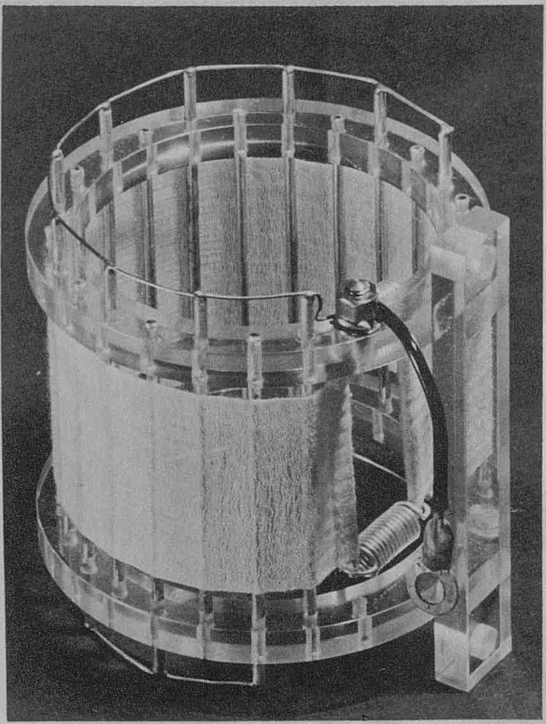


FIG. 8. GREGORY HUMIDIOMETER ELEMENT.

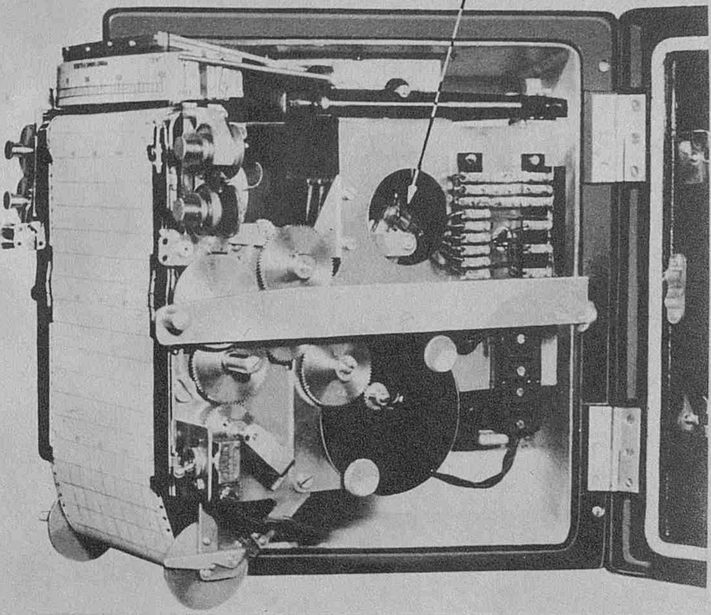

Mercury Switches.

FIG. 6. NEGRETTI & ZAMBRA RECORDER.

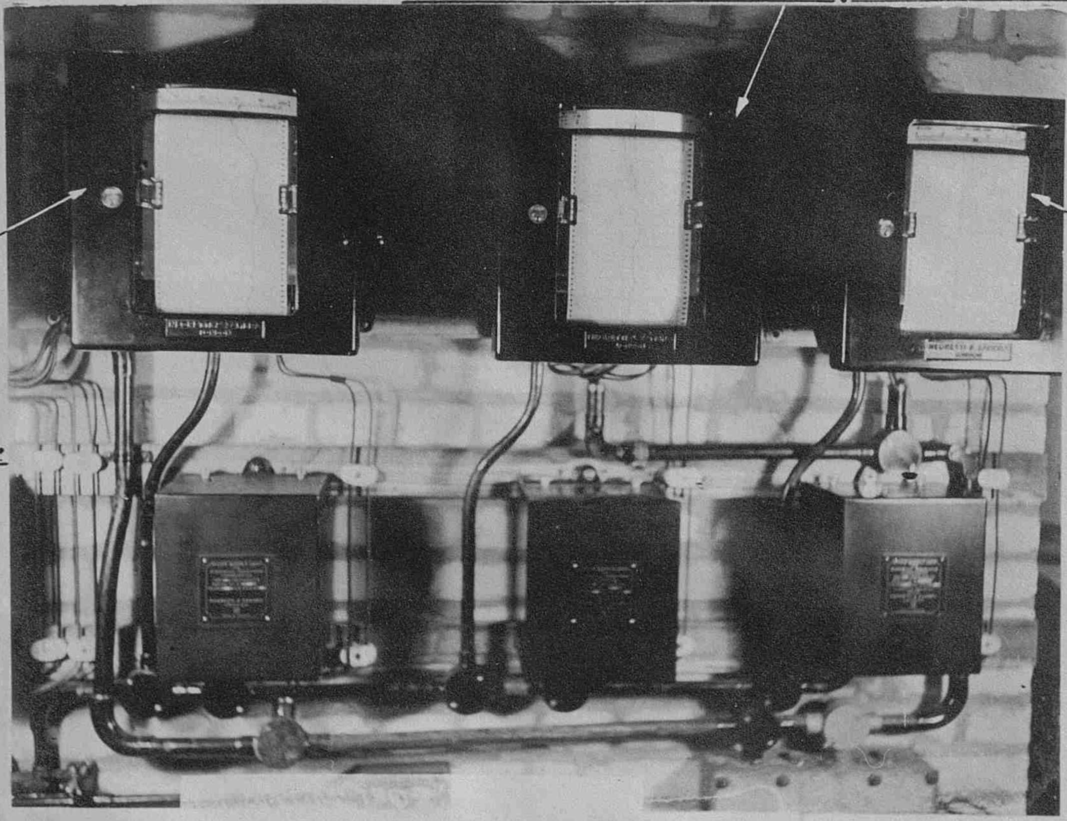
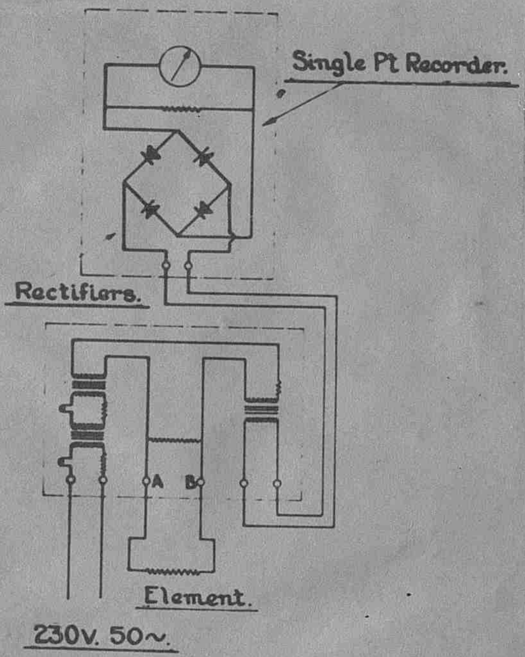

Differential Temperature.
Single Point Humidity.
Single Point Temperature

FIG. 7. RECORDERS MOUNTED ON WALL.


Single Pt Recorder.
Rectifiers.
A
B
Element.
230V. 50~.

The above circuit is for the single p't humidity recorder. For the 4 p't instrument the only difference is that A B, the terminals to the external circuit, are connected via mercury switches to separate elements as shown in Fig.11.

Fig 10. Measurement of Relative Humidity.

HEN 1·9·44

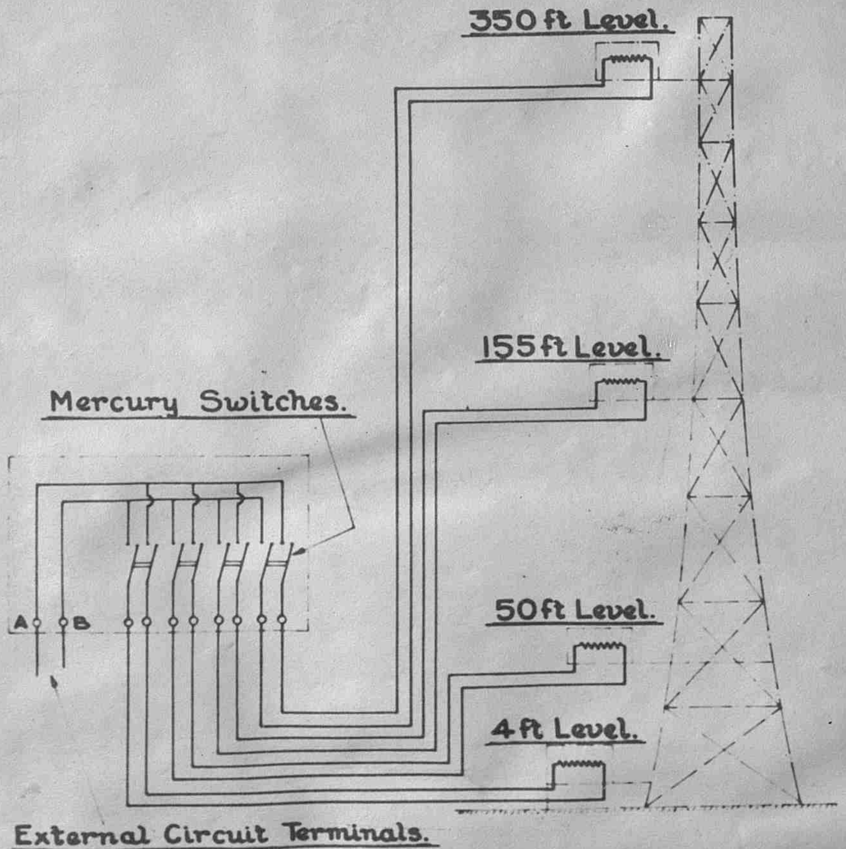
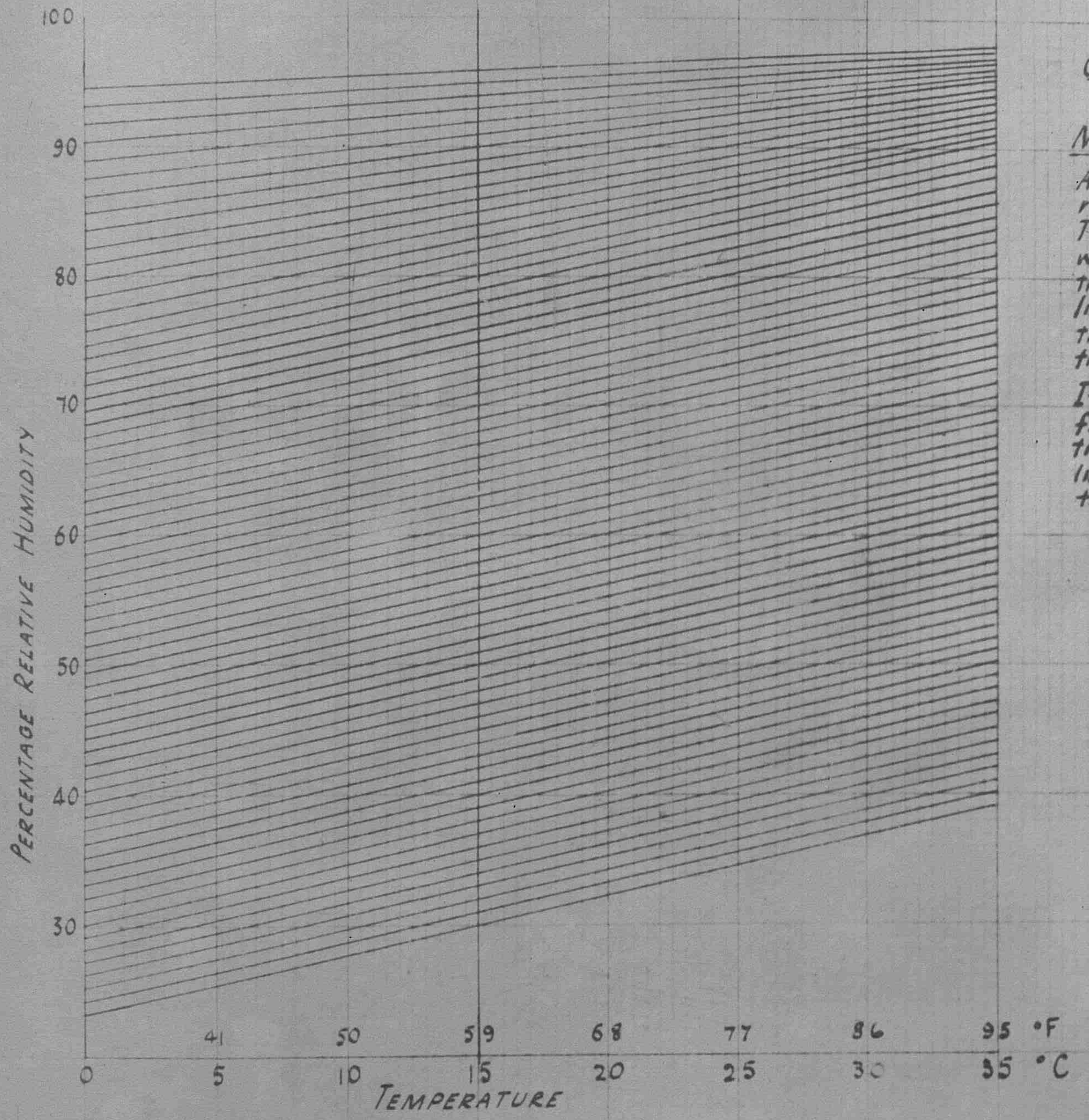

350 ft Level.
155 ft Level.
Mercury Switches.
A
B
50 ft Level.
4 ft Level.
External Circuit Terminals.

Fig 11. External Circuit for 4pt., Humidity Recorder.

DEH 2·8·44

NEGRETTI & ZAMBRA



Calibration Point 15°C.

Method of use.

As an example let the instrument reading be 50% R.H. and the Temperature 30·5C. Find the point where the 50% R.H. ordinate intersects the 30·5°C ordinate. Follow down the line from this point. Where it meets the 15°C line, i.e. 43% R.H. is the true Humidity.

If the readings are 50% R.H. and 6°C from where these ordinates intersect trace line upwards till it meets the 15°C line, i.e. at 54% R.H. which is the true Humidity.

FIG. 9.

TEMPERATURE CORRECTION TABLE.

A.M. T.R.E.

W.H. 8811

DRWG. NO. H.O. 1168-B

DATE: 13·10·43
SC: —
DR:
TR:
CH:

122 REGENT STREET LONDON W.1.

METEOROLOGICAL OFFICE, AIR MINISTRY

Synoptic Divisions Technical Memorandum No. 102

CONVECTION
IN THEORY AND PRACTICE

BY

SVERRE PETTERSSEN,
E. KNIGHTING,
R. W. JAMES, and
N. HERLOFSON

Acc. 47.566

March, 1945

(71703)

CONVECTION IN THEORY AND PRACTICE

CONTENTS

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3. The parcel method : Normand's classification.
4. The parcel method : cloud base, top and amount.
5. The parcel method : magnitude of velocities and accelerations.
6. The slice method : stability criteria.
7. The slice method : the positive area.
8. Further remarks on the vertical velocity.
9. The slice method : cloud base, top and amount.
10. The convection temperature.
11. The use of virtual temperature.
12. Remarks on observational accuracy.
13. Release of precipitation.

III. STATISTICAL CHARACTERISTICS

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15. Characteristics of selected situations.
16. Remarks on cloud observations.
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24. Release of precipitation.
25. The lapse rate within convective clouds.
26. Vertical velocity and gusts.
27. Relation to synoptic situation.
28. Some cases of deep convection.

REFERENCES TO LITERATURE

SUMMARY

A critical summary of theories of convection, notably those which lead to the so-called parcel method and slice method of investigating the stability conditions, is given in Section II.

Expressions for the amount of energy, the vertical velocity and acceleration are derived, and the results are compared with Normand's classification of ascent curves. It is shown (1) that both methods lead to consistent criteria for the release of convection from the ground, (2) that the parcel method underestimates the amount of available energy when both the ascending and the descending mass follow the same adiabatic, and (3) that the parcel method greatly overestimates the amount of available energy when saturated air ascends through a dry environment.

The slice method leads to the result that the amount of energy available for the creation of convective currents, depends not only upon the lapse rate of the stratification, but also upon the kinematics of the motion. The slice method, although theoretically sound, is difficult to apply in practice, and in most cases, only qualitative results can be obtained.

The findings of the slice method and of the parcel method are used to obtain assessments of the amount of convective cloud and the height to which the convective cloud will reach. Except when the convection is limited to a shallow layer (i.e. cumulus humulis and strato-cumulus) the two methods give widely different results. Methods of forecasting the convection temperature, the maximum temperature, and the condensation level are discussed in connection with the effective amounts of energy derived by Gold. The use of virtual temperature in the analysis of instability phenomena is discussed, and it is shown that virtual temperature and virtual saturated-adiabatics should be used instead of temperature and ordinary saturated-adiabatics. The influence of observational errors is discussed, and the conditions for release of precipitation from convective clouds are summarised.

The theoretical relationships have been compared with actual observations, and the results are discussed in Section III. It is shown that (1) both the parcel method and the slice method give satisfactory criteria to account for the occurrence of convective cloud, (2) while the parcel method gives no information as to the amount of cloud, the slice method gives results that are largely in agreement with observations, (3) the base of convective cloud is about 25 mb. above the condensation level evaluated from simultaneous observations, but the forecast condensation level is in poor agreement with what is observed, (4) the slice method gives satisfactory results for the heights to which the tops of the convective cloud will reach, whereas the parcel method greatly overestimates these heights.

The release of convective currents from the ground is discussed in considerable detail, and it is shown that convective clouds are likely to appear as soon as the stability of the surface layer is reduced so much that the negative area, as measured on the TePhi-gram, is $0 \cdot 5$ cm^2., or less. The release of convection caused by diurnal heating is discussed, and it is shown that Gold's method of evaluating the maximum temperature is largely satisfactory, and so is also the method of evaluating the convection temperature, except when the structure of the surface layer is complex.

The equivalent amount of cloud is introduced to take account of the horizontal as well as the vertical extent of the clouds, and it is shown that this amount is, on the whole, proportional to the available energy.

The probability of release of precipitation is found to increase (*a*) with the temperature below freezing at the top of the cloud, and (*b*) with the depth of the cloud.

The lapse rate of temperature within convective cloud is tabulated and discussed in connection with gusts and accelerations. Although the accelerations are very small, vertical gusts in the vicinity of 20 m/sec. are likely to be found in the most active cumulo-nimbus.

The occurrence of convection has been correlated with the vertical wind shear and the vorticity at the 750 mb. level. The observations show that while shallow convection is unrelated to the vorticity, there is a very high association between deep convection and cyclonic vorticity.

Some typical cases of deep convection are tabulated, and their characteristics discussed.

I. INTRODUCTION

The vertical currents in the atmosphere may be classified according to their horizontal dimensions, viz. :—

(1) *Turbulent eddies*, whose dimensions are of the order of magnitude of metres or tens of metres, and whose life is so short that the individual eddies do not normally become visible through the formation of clouds.

(2) *Convective currents*, whose dimensions are of the order of hundreds or thousands of metres, and whose life is of the order of a few hours. These currents will normally become visible through the formation of clouds in their ascending branches.

(3) *Large scale circulations*, whose dimensions are of the order of hundreds or thousands of kilometres (e.g. extra-tropical cyclones and anticyclones), and whose life is of the order of days. These circulations readily become visible through the formation of extensive cloud systems in their ascending branches.

The above classification does not represent a grouping according to arbitrary limits of what is small, medium, or large, for it is not observed that a turbulent eddy assumes the dimensions and characteristics of a cumulo-nimbus cloud, nor does a cumulo-nimbus cloud ever approach the dimensions of an extra-tropical cyclone. Thus, as far as horizontal dimensions are concerned, the convective currents stand out as a distinct type.

The vertical velocities associated with the above-mentioned currents are equally distinct. In the large scale circulations the vertical velocities are of the order of a few centimetres per second, whereas those associated with the convective currents are of the order of a few metres per second. Observations show that the vertical velocities in the convective currents are not likely to exceed 10–15 m/sec. Any satisfactory theory of convection must give vertical velocities that do not greatly exceed this limit.

Amongst the numerous contributions to the theory of convection and to the technique of forecasting convective phenomena, only those which have a direct bearing on the present investigation will be mentioned here.

The first contribution to the theory of convection was rendered by Lord Kelvin (1865) who, by considering the compressibility of dry air, derived the expression for the dry-adiabatic lapse rate. The next principal contribution is due to Hertz (1884) who, in addition to the compressibility, considered the liberation of the latent heat of vaporisation, and developed an adiabatic chart with dry and saturated adiabatics; a diagram that was later perfected and introduced into synoptic practice by Stüve (1922).

Already during the later part of the First World War, Douglas (1920) studied convective phenomena, notably thunderstorms, and correlated their occurrence with the synoptic situations. Douglas' empirical results are of considerable interest, particularly since they show that the occurrence of thunderstorms is definitely related to the general synoptic situation.

Later investigations of convection have dealt mainly with the thermodynamical relationships, the criteria for static stability and instability, and the amount of energy available for the creation of convective currents. These investigations have resulted in two different methods of investigating convective phenomena, and two sets of criteria for stable and unstable stratifications. We shall refer to these methods as the Parcel Method and the Slice Method.

The parcel method, which is mainly due to Refsdal (1930) and Normand (1938), results from considerations of the buoyancy forces acting on a parcel of air that is displaced adiabatically in an undisturbed environment. The stability criteria thus obtained can be expressed in terms of lapse rate only, and the amount of energy available for the creation of convective currents can be obtained from a plot on an indicator diagram (e.g. a TePhi-gram).

The slice method, which was introduced by Bjerknes (1938) and extended by Petterssen (1939), is derived by considering the downward as well as the upward currents, having regard to the equation of continuity. The stability criteria thus derived, depend not only upon the lapse rate but also upon the kinematics of the motion. The expressions for the amounts of energy derived by the aid of the slice method differ essentially from those derived by the aid of the parcel method, the difference between the two amounts being proportional to the ratio of the ascending mass to the descending mass.

The essential difference between the parcel method and the slice method is that the adiabatic heating in the descending currents is neglected in the former and accounted for in the latter, with the result that in all cloud-producing processes the parcel method renders amounts of energy, vertical accelerations and velocities that are much too large. The results derived by the aid of the slice method converge towards those derived by the aid of the parcel method when the horizontal dimensions of the ascending mass are small as compared with those of the descending mass.

The difficulty in the way of using the slice method in practice is due to the circumstance that the stability criteria and the amounts of available energy depend upon the ratio of the ascending mass to the descending mass, and since this ratio is not observed, only qualitative results can be obtained.

Regarding convection from a forecasting point of view, one would naturally desire criteria for : the commencement and cessation of convective currents, the vertical velocities and gusts, the vertical accelerations, the heights of the base and of the top of the convective clouds, the cloud amount, and the amount of precipitation. The aim of this investigation is to determine the practical applicability of the theories of convection to the analysis and forecasting of the said phenomena.

Section II of this report contains a summary and, in certain respects, an extension of the present theories with a formulation suitable for comparison with actual observations. About 300 aerological ascents and observations of clouds in convective situations have been analysed and compared with the theoretical relationships. The results are described in Section III, which also contains some statistical results relating the occurrence of convection to the synoptic situation.

II. SOME THEORETICAL RELATIONSHIPS

1. The Parcel Method : Stability Criteria

The parcel method of investigating the stability of a stratification is based upon considerations of a parcel of air that is displaced adiabatically from its original position in such a manner that the environment remains undisturbed.* The stratification is said to be stable, or unstable, according to whether the parcel becomes accelerated in the direction towards, or away from, the level whence it came.

We consider a parcel of air embedded in an environment that is in hydrostatic equilibrium, and let letters with indices refer to the parcel and letters without indices refer to the environment. Then :

$$\dot{w}' = -\alpha' \frac{\partial p'}{\partial z} - g$$

$$o = -\alpha \frac{\partial p}{\partial z} - g \qquad \dots \quad \dots \quad \dots \quad \dots \quad \dots \qquad (1)$$

Here,

$\dot{w}$ = vertical acceleration
α = specific volume
p = pressure
g = acceleration of gravity
z = distance along a vertical axis.

Since the pressure gradient along the vertical is determined by the surrounding air, we have

$$\frac{\partial p'}{\partial z} = \frac{\partial p}{\partial z}$$

and therefore,

$$\dot{w}' = g \frac{\alpha' - \alpha}{\alpha} \qquad \dots \quad \dots \quad \dots \quad \dots \quad \dots \qquad (2)$$

or, from the equation of state

$$\dot{w}' = g \frac{T' - T}{T} \qquad \dots \quad \dots \quad \dots \quad \dots \quad \dots \qquad (3)$$

where T = absolute temperature.

It will thus be seen that the parcel will be accelerated upwards if it is warmer, and downwards if it is colder, than the environment.

It is of interest to note that the only assumption made in the derivation of eq. (3) is that the air surrounding the parcel is not accelerated along the vertical ; no assumption is made as regards adiabatic processes and vertical velocities.

If the horizontal distribution of temperature could be observed in sufficient detail, the static stability could be evaluated by the application of eq. (3). However, since the distance between the observing stations is large whereas the ascents give a continuous record of the temperature distribution along the vertical, it is more convenient to express the stability criteria in terms of the lapse rate of temperature.

We consider next a parcel of air at an arbitrary level z_0 where the temperature is T_0 and the lapse rate γ. Then

$$T = T_0 - \gamma \Delta z$$

describes the temperature of the environment around the point considered. If the parcel is displaced adiabatically a small distance Δz, its temperature will vary according to the adiabatic law, *viz.*,

$$T = T_0 - \gamma_a \Delta z$$

If, during this displacement, the temperature of the environment remains unchanged,

$$T' - T = (\gamma - \gamma_a) \Delta z$$

and eq. (3) becomes

$$\dot{w}' = g \frac{\gamma - \gamma_a}{T} \Delta z \qquad \dots \quad \dots \quad \dots \quad \dots \quad \dots \qquad (4)$$

which gives the stability criteria of the parcel method. It will be seen that the acceleration has the same sign as the displacement (i.e. instability) when

$$\gamma > \gamma_a$$

and its sign is opposite to that of the displacement (i.e. stability) when

$$\gamma < \gamma_a$$

Accordingly, the parcel method leads to the following criteria for stability and instability :

stable equilibrium : $\gamma < \gamma_a$
indifferent equilibrium : $\gamma = \gamma_a$
unstable equilibrium : $\gamma > \gamma_a$

In the above expressions, the dry-adiabatic (γ_d) or the saturated-adiabatic (γ_s) lapse rate should be used according to whether the parcel is non-saturated or saturated with moisture.

Although the equations (3) and (4) are similar in structure, there is a fundamental difference between the assumptions underlying them.

Eq. (3) was derived on the sole assumption that the vertical acceleration in the environment is nil, or negligible. No assumption was made as regards the vertical velocity, nor were adiabatic changes introduced. The equation relates the excess temperature of the parcel to the acceleration relative to the environment, but it does not indicate how the excess temperature was established.

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* The parcel method has recently been extended by Sir Charles B. Normand and his associates in India, by making plausible assumptions as to the efficiency of the overturning processes. The results of these investigations are as yet unpublished, and the following discussion deals with the theory and practice of the parcel method as presented in standard texts.

Eq. (4), on the other hand, was derived from eq. (3) on the basis of the additional assumptions that the motion of the parcel is adiabatic and that there is no vertical velocity in the environment. This latter assumption represents a severe restriction. We shall return to the discussion of this in a later paragraph. Here, it suffices to remark that eq. (4), and not eq. (3), expresses the contents of the parcel method. Eq. (4), however, represents a special case of eq. (3).

2. **The Parcel Method : Energy, Velocity and Acceleration**

The theory underlying the parcel method can be extended to give some simple relationships between the vertical accelerations and velocities and the areas obtained from the plots on a TePhi-gram, or other diagrams which represent an equal-area transformation of the pressure-volume diagram. The following discussion will refer to the TePhi-gram (Form 2810 and 2810A).

The vertical velocity of the parcel may be written

$$w' = \frac{dz}{dt}$$

and the acceleration

$$\dot{w}' = \frac{dw'}{dt}$$

Substitution into eq. (3) gives

$$\frac{d(\tfrac{1}{2}w'^2)}{d(gz)} = \frac{T' - T}{T} \qquad \text{.. .. ..} \qquad (5)$$

Since (T′ − T) is very small as compared with T, it follows that the variation in the kinetic energy is only a small fraction of the variation in the potential energy of the parcel.

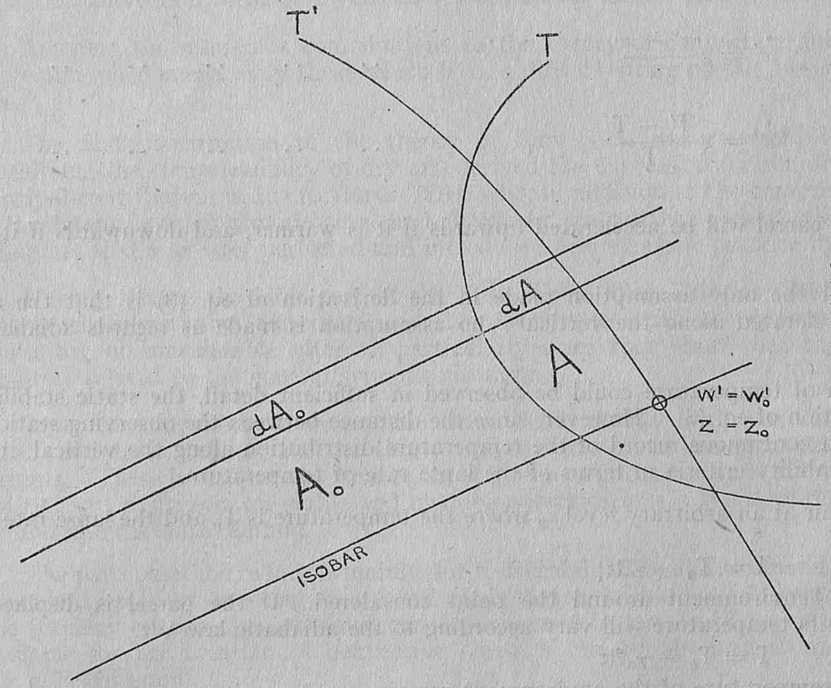

Fig. 1.—Schematical section of the TePhi-gram. T′ = temperature of the parcel; T = temperature of the environment; The area A_0 extends to the point where the Absolute temperature is zero.

We consider next Fig. 1 where T refers to the environment and T′ to the parcel. Let A_0 denote the area between two isobars to the left of the curve T, and let A denote the area bounded by the same isobars and the curves T and T′. The TePhi-gram being an equal-area transformation of the pressure-volume diagram,

$$d(gz) = k\,dA_0 \qquad \text{.. .. ..} \qquad (6)$$

where *k* is the scale factor of the TePhi-gram relating area to energy.

On the TePhi-gram, the distance along an isotherm from one isobar to another is independent of the temperature, and from this it follows that

$$dA = dA_0 \frac{T' - T}{T} \qquad \text{.. .. ..} \qquad (7)$$

which combined with eqs. (5) and (6) gives

$$d(\tfrac{1}{2}w'^2) = k\,dA$$

and therefore

$$\tfrac{1}{2}w'^2 - \tfrac{1}{2}w_0'^2 = kA \qquad \text{.. .. ..} \qquad (8)$$

where A is measured from the isobar where the parcel was at the initial moment. It will be seen from eq. (7) that *d*A is positive when the parcel is warmer, and negative when the parcel is colder, than the environment.

Solving with respect to the vertical velocity, we obtain :

$$w' = \sqrt{2(\tfrac{1}{2}w_0'^2 + kA)}$$

Now, in the TePhi-gram referred to above

$$k = \frac{1}{36}\ \frac{\text{Joule}}{\text{gm cm}^2}$$

and, when A is expressed in cm.² and the velocity in m/sec.

$$w' = \sqrt{w_0'^2 + \frac{1000}{18}A} \qquad \text{.. .. ..} \qquad (9)$$

If the vertical velocity is zero at the initial moment

$$w' = 7{\cdot}45\sqrt{\mathrm{A}} \quad \ldots \quad (10)$$

or

$$\mathrm{A} = 0{\cdot}018\, w'^2 \quad \ldots \quad (11)$$

On the other hand, if the initial velocity of the parcel is w'_0, it can describe the negative area

$$\mathrm{A} = -\,0{\cdot}018\, w_0'^2 \quad \ldots \quad (12)$$

before its kinetic energy is consumed.

We shall return to the discussion of these relationships in connection with the discussion of the statistical results in Section III.

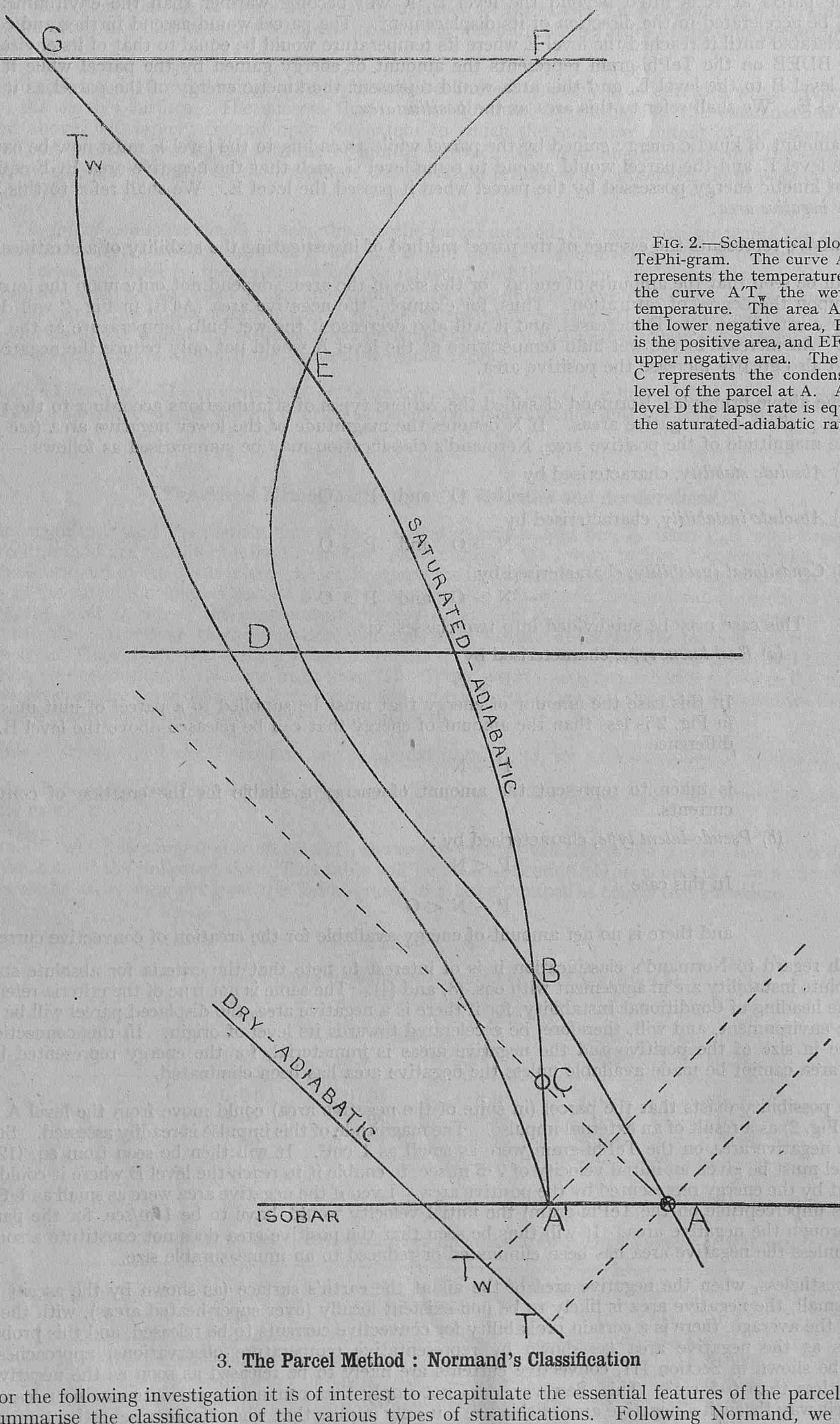

FIG. 2.—Schematical plot on a TePhi-gram. The curve ABEF represents the temperature, and the curve A′T_w the wet-bulb temperature. The area ACB is the lower negative area, BDEF is the positive area, and EFG the upper negative area. The point C represents the condensation level of the parcel at A. At the level D the lapse rate is equal to the saturated-adiabatic rate.

3. The Parcel Method : Normand's Classification

For the following investigation it is of interest to recapitulate the essential features of the parcel method and summarise the classification of the various types of stratifications. Following Normand, we consider a TePhi-gram in which are plotted the ascent curves of temperature T and wet-bulb temperature T_w, as shown diagrammatically in Fig. 2.

To illustrate the application of the parcel method to actual ascents, we consider a parcel of air at an arbitrary level, say A, where the temperature is T and the wet-bulb temperature is T_w. If the parcel is displaced upwards, it will cool dry-adiabatically and become saturated at the level C, whence it cools at the saturated-adiabatic rate.

During its ascending motion, from A to B, the parcel will be colder than the environment and, therefore, be accelerated downwards, i.e. against its displacement. The area ABC in Fig. 2 represents the amount of energy that must be supplied to a parcel of unit mass at A in order to overcome the buoyancy forces and bring the parcel to the level B. We shall refer to this area as the *negative area.*

If the parcel at A is lifted beyond the level B, it will become warmer than the environment and, therefore, be accelerated in the direction of its displacement. The parcel would ascend further and continue to be accelerated until it reached the level E where its temperature would be equal to that of its environment. The area BDEB on the TePhi-gram represents the amount of energy gained by the parcel while it moves from the level B to the level E, and this area would represent the kinetic energy of the parcel as it arrives at the level E. We shall refer to this area as the *positive area.*

The amount of kinetic energy gained by the parcel while ascending to the level E must now be expended above the level E, and the parcel would ascend to some level G, such that the negative area EGF equals the amount of kinetic energy possessed by the parcel when it passed the level E. We shall refer to this area as the *upper negative area.*

The above represents the essence of the parcel method of investigating the stability of a stratification.

It will be seen that the amounts of energy, or the size of the areas, depend not only upon the lapse rates but also upon the degree of saturation. Thus, for example, the negative area (ACB) in Fig. 2 will decrease as the lapse rate from A to B increases, and it will also decrease if the wet-bulb temperature at the level A increases. An increase in the wet-bulb temperature at the level A would not only reduce the negative area but would also greatly increase the positive area.

On the above principles, Normand classified the various types of stratifications according to the relative size of the positive and negative areas. If N denotes the magnitude of the lower negative area (see Fig. 2) and P the magnitude of the positive area, Normand's classification may be summarised as follows :—

(1) *Absolute stability*, characterised by

$$N > O \quad \text{and} \quad P = O$$

(2) *Absolute instability*, characterised by

$$N = O \quad \text{and} \quad P > O$$

(3) *Conditional instability*, characterised by

$$N > O \quad \text{and} \quad P > O$$

This case may be subdivided into two classes, viz.,

(*a*) *Real latent type*, characterised by

$$P > N$$

In this case the amount of energy that must be supplied to a parcel of unit mass at A in Fig. 2 is less than the amount of energy that can be released above the level B. The difference

$$P - N$$

is taken to represent the amount of energy available for the creation of convective currents.

(*b*) *Pseudo-latent type*, characterised by :

$$P < N$$

In this case

$$P - N < O$$

and there is no net amount of energy available for the creation of convective currents.

With regard to Normand's classification it is of interest to note that the criteria for absolute stability and absolute instability are in agreement with eqs. (3) and (4). The same is not true of the criteria referred to under the heading of Conditional Instability, for if there is a negative area, the displaced parcel will be colder than the environment, and will, therefore, be accelerated towards its level of origin. In this connection the difference in size of the positive and the negative areas is immaterial, for the energy represented by the positive area cannot be made available unless the negative area has been eliminated.

The possibility exists that the parcel (in spite of the negative area) could move from the level A to the level B (Fig. 2) as a result of an external impulse. The magnitude of this impulse is readily assessed. Suppose that the negative area on the TePhi-gram were as small as 1 cm^2. It will then be seen from eq. (12) that the parcel must be given an initial velocity of 7·5 m/sec. to enable it to reach the level B where it could begin to benefit by the energy represented by the positive area. Even if the negative area were as small as 1·8 mm^2. (which is imperceptible on the TePhi-gram) the initial velocity would have to be 1 m/sec. for the parcel to move through the negative area. It will thus be seen that the positive area does not constitute a source of energy unless the negative area has been eliminated or reduced to an immeasurable size.

Nevertheless, when the negative area of the air at the earth's surface (as shown by the ascent curve) is very small, the negative area is likely to be non-existent locally (over super-heated areas), with the result that, on the average, there is a certain probability for convective currents to be released, and this probability increases as the negative area (as shown by representative temperature observations) approaches zero. As will be shown in Section III, convective currents are likely to be released as soon as the negative area becomes less than 0·5 cm^2. This, however, does not imply that convective currents can be released and force their way through a negative area of this size ; it only implies that when the representative temperature curve shows a negative area less than 0·5 cm.^2, the area is likely to be absent locally. As the heating continues, the negative area disappears generally, and convection becomes widespread.

4. The Parcel Method : Cloud Base, Top and Amount

The assumptions underlying the parcel method lead to some definite relations between the ascent curve and the levels at which the base and the top of the convective cloud should be found.

(1) *The condensation level* is obtained from the ascent curve as follows. Let A in Fig. 2 represent the temperature and A′ the wet-bulb temperature of a parcel of air near the earth's surface. If the parcel is lifted adiabatically with constant moisture content, its temperature point will follow the dry-adiabatic through A, and its wet-bulb temperature point will follow the saturated-adiabatic through A′. The condensation level is then the point of intersection of these two adiabatics. The condensation level, thus determined, refers to convection that starts from the earth's surface. In rare cases,* convection may be released from some higher level, in which case the temperature and the wet-bulb temperature of the appropriate level must be considered.

The discussion in the following paragraphs will be limited to convective currents which commence at, or near, the earth's surface. The success that can be obtained by evaluating the condensation level, as described above, will largely depend upon the extent to which the moisture content of the ascending air is conserved, or, in other words, the amount of mixing between the ascending currents and the environment. A statistical comparison between the computed condensation level and the observed cloud base will be given in Section III.

(2) *The top of convective clouds.*—According to the parcel method, the ascending air would rise on account of the buoyancy forces above the level B in Fig. 2, and arrive at the level E with a maximum of kinetic energy. Above the level E, the motion would be retarded and the parcel would continue to the top of the upper negative area, i.e. to G in Fig. 2. Accordingly, the tops of the convective clouds would be observed at, or near, the top of the upper negative area. As will be shown in Section III, the observed cloud tops rarely reach as far as to the top of the positive area, showing that there is a vast difference between the computed and the observed heights to which the convective clouds reach.

(3) *Cloud amount.*—The environment being undisturbed, the parcel method renders no upper limit to the amounts of convective cloud, and no relation between the cloud amount and the ascent curve.

5. The Parcel Method : Magnitude of Velocities and Accelerations

The magnitude and the distribution of the vertical velocities and accelerations that would result from the parcel method are readily obtained from the ascent curves. We return to Fig. 2 and consider a parcel of air at the lower end of the positive area, i.e. at B, where the initial vertical velocity is supposed to be negligible. As soon as the parcel is displaced beyond this level, acceleration sets in, and the acceleration reaches a maximum value at the level D, where the temperature difference between the parcel and the environment reaches a maximum value. However, the acceleration continues until the parcel reaches the level E, i.e. the top of the positive area. Here, then, the vertical velocity reaches its highest value. These conclusions will be discussed in relation to observational evidence in Section III. What we are here concerned to show is the numerical relations between the areas obtained from the analysis of the TePhi-gram and the vertical velocities that would be observed if the parcel method were reasonably accurate.

Table A shows the vertical accelerations computed from eq. (3) for various values of T′ and T.

Table B is computed from eq. (4) for various deviations of the actual lapse rate from the appropriate adiabatic rate.

Table C, which is computed from eq. (11), shows the vertical velocity of the parcel after it has described a positive area of the indicated size. This table will be used in Section III in connection with the statistical analysis of the areas obtained from the TePhi-grams in a large number of convective situations.

TABLE A

Vertical Accelerations (*m/sec.*^{2}) *computed from Equation* 3

T ° C.	T′—T (° C.)				
	±1	±5	±10	±15	±20
−30	±0·04	±0·20	±0·40	±0·60	±0·81
0	±0·04	±0·18	±0·36	±0·54	±0·72
+30	±0·03	±0·16	±0·32	±0·49	±0·65

TABLE B

Vertical Accelerations (*m/sec.*^{2}) *computed from Equation* 4 *for a Displacement of* 100 *metres*

T ° C.	γ—γ_a (° C. per 100 m.)				
	−3	−2	−1	0	0·5
−30	−0·12	−0·08	−0·04	0	0·020
0	−0·11	−0·07	−0·04	0	0·018
+30	−0·10	−0·06	−0·03	0	0·016

---

* The qualification " rare " refers to the observational material examined. In tropical and sub-tropical regions, convective instability is often present, and convection is released aloft.

TABLE C

Vertical Velocities (m/sec.) Corresponding to Areas (cm²) on the TePhi-gram, computed from Equation 11

Vertical Velocity m/sec.	Area cm²
1	0·02
2	0·07
5	0·45
10	1·80
15	4·05
20	7·20
30	16·20
40	28·80
50	45·00

In connection with Table C it is of interest to note that the most frequent size of the positive area in convective situations is about 15 cm² to which corresponds, according to the parcel method, a vertical velocity at the top of the positive area of about 30 m/sec. In extreme cases, positive areas as large as 60 cm² have been observed, to which would correspond a vertical velocity of about 60 m/sec.

6. **The Slice Method : Stability Criteria**

In the foregoing paragraphs we considered a parcel of air moving in an undisturbed environment, and it was shown that the assumptions of an undisturbed environment lead to values of the vertical velocity and the vertical acceleration that are vastly in excess of what is believed to occur in the atmosphere.

In the slice method, as developed by Bjerknes (1938) and Petterssen (1939), a horizontal slice of air of unit thickness is considered. Let M′ be the mass that moves upwards and M the mass that moves downwards through the slice, and let the corresponding velocities be w' and w. On the assumption that the slice is so large that the transport of mass through the vertical walls is negligible in comparison with the transport through the horizontal surfaces, there must be as much mass moving upwards as downwards through the slice, or

$$M'w' = -Mw \qquad \text{.. .. ..} \qquad (13)$$

The theory underlying the slice method has been described in detail elsewhere (Petterssen 1940), but a simplified derivation of the essentials will be given here in order to show more precisely the points where the slice method renders results that differ from those obtained by the parcel method. In particular, we shall be concerned with the relationship between the areas on the TePhi-gram and the vertical velocity and acceleration.

We consider a schematical TePhi-gram as shown in Fig. 3.

Let the line AB represent the ascent curve at the initial moment. We consider the point A on the ascent curve (whose pressure is p_0) and assume that the air at this point is saturated. If this air ascends, it will move along the saturated adiabatic and arrive at the point C on the isobar p. To maintain continuity, a certain mass must descend, and if the descending air is non-saturated (as it would be outside a cloud), it will move along a dry-adiabatic. The following is readily seen from Fig. 3.

(1) The condition for the ascending air to arrive at the isobar p without any excess temperature is that the descending air at that level originated from the level B.

(2) The condition for the ascending air to arrive at the isobar p with a temperature in excess of its environment is that the descending air originated at some level lower than B.

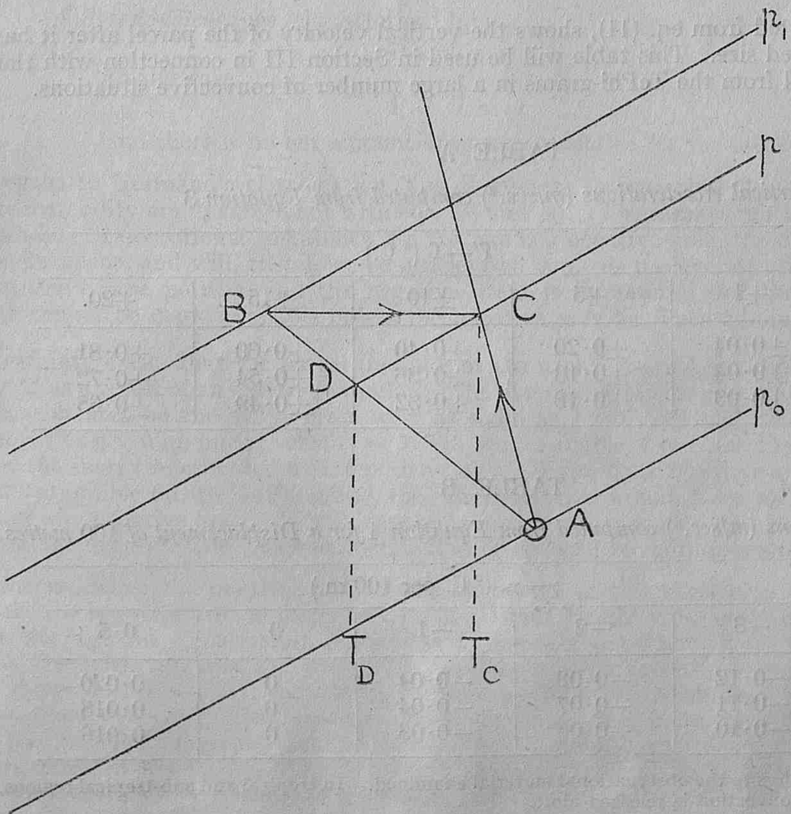

FIG. 3.—Schematical section of a TePhi-gram. AB = temperature curve; AC = saturated-adiabatic; BC = dry-adiabatic. According to the parcel method, the excess temperature is $T_C - T_D$, and the acceleration would be proportional to the excess temperature. According to the slice method, the temperature of the environment would increase by the amount BC while the parcel moves from A to C.

(3) The condition for the ascending air to arrive at the isobar p with a temperature lower than that of its environment is that the descending air originated at some level higher than B.

(4) If the environment remained undisturbed (as is assumed in the parcel method), the excess temperature of the ascending air at C would be

$$\varDelta \mathrm{T} = \mathrm{T_C} - \mathrm{T_D} = (\gamma - \gamma_s)\varDelta z$$

It will thus be seen that, in the case illustrated in Fig. 3, the parcel method gives an excess temperature of the ascending air that is too high, and, therefore, an acceleration that is too large.

On the above principles, it is easy to obtain a general formula for the excess temperature of the ascending air. Let γ be the lapse rate as shown by the ascent curve in Fig. 3, $\varGamma$ the adiabatic lapse rate of the descending air, and $\varGamma'$ the adiabatic lapse rate of the ascending air. (If both masses are either saturated or non-saturated $\varGamma = \varGamma'$.)

The excess temperature of the ascending air after a time interval t is then

$$\begin{aligned} \varDelta \mathrm{T} = \mathrm{T}' - \mathrm{T} &= \mathrm{T_A} - \varGamma' w' t - (\mathrm{T_B} - \varGamma w t) \\ &= \mathrm{T_A} - \mathrm{T_B} + (-\varGamma' w' + \varGamma w)t \\ &= -\gamma(w - w')t + (-\varGamma' w' + \varGamma w)t \end{aligned}$$

and from eq. (13)

$$\varDelta \mathrm{T} = w't\left[\gamma - \varGamma' - \frac{\mathrm{M}'}{\mathrm{M}}(\varGamma - \gamma)\right] \quad \ldots \quad (14)$$

or since

$$w't = \varDelta z$$

$$\varDelta \mathrm{T} = \varDelta z\left[\gamma - \varGamma' - \frac{\mathrm{M}'}{\mathrm{M}}(\varGamma - \gamma)\right] \quad \ldots \quad (15)$$

and also from eq. (13)

$$\varDelta \mathrm{T} = \varDelta z\left[\gamma - \varGamma' + \frac{w}{w'}(\varGamma - \gamma)\right] \quad \ldots \quad (16)$$

In the above formulae, w is by definition negative and w' positive.

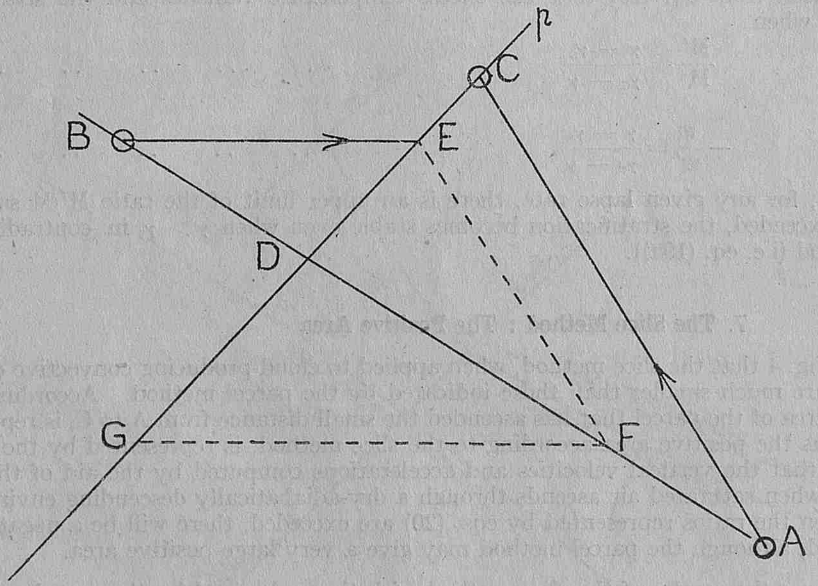

FIG. 4.—Schematical section of a TePhi-gram. AB = temperature curve; AC = adiabatic of the ascending air; BE = adiabatic of the descending air; GC = isobar. The excess temperature is proportional to EC. The ascending velocity proportional to AC, and the descending velocity is proportional to BE.

The geometrical interpretation of the above equation is readily obtained from Fig. 4. Thus, if the air at A ascends to C and the air at B descends to E in the time interval t, we have for the excess temperature of the ascending air when it arrives at the isobar p:

$$\begin{aligned} \frac{\varDelta \mathrm{T}}{(\gamma - \varGamma')\, w't} &= \frac{\mathrm{EC}}{\mathrm{DC}} = 1 - \frac{\mathrm{DE}}{\mathrm{DC}} = 1 - \frac{\mathrm{DF}}{\mathrm{DA}} \\ &= 1 - \frac{\mathrm{DF}}{\mathrm{DB}} \cdot \frac{\mathrm{DB}}{\mathrm{DA}} = 1 - \frac{\mathrm{DG}}{\mathrm{DE}} \cdot \frac{\mathrm{DB}}{\mathrm{DA}} \\ &= 1 - \frac{\varGamma - \gamma}{\gamma - \varGamma'} \cdot \left(-\frac{w}{w'}\right) = 1 - \frac{\varGamma - \gamma}{\gamma - \varGamma'}\,\frac{\mathrm{M}'}{\mathrm{M}} \end{aligned}$$

or, since

$$w't = \varDelta z,$$

$$\varDelta \mathrm{T} = \varDelta z\left[\gamma - \varGamma' - \frac{\mathrm{M}'}{\mathrm{M}}(\varGamma - \gamma)\right]$$

which is identical with eq. (16).

Eq. (16) can now be specialised as follows:

(1) If both the ascending and the descending masses are non-saturated (e.g. under the cloud base)

$$\varGamma = \varGamma' = \gamma_{\mathrm{d}}$$

and the excess temperature of the ascending air is

$$\varDelta \mathrm{T} = \varDelta z\left(1 + \frac{\mathrm{M}'}{\mathrm{M}}\right)(\gamma - \gamma_{\mathrm{d}}) = \varDelta z\left(1 - \frac{w}{w'}\right)(\gamma - \gamma_{\mathrm{d}}) \quad \ldots \quad (17)$$

In the parcel method $w = o$, and therefore

$$\varDelta \mathrm{T} = \varDelta z\,(\gamma - \gamma_{\mathrm{d}}) \quad \ldots \quad (17a)$$

It will thus be seen that *the parcel method and the slice method give the same sign of the excess temperature, but the parcel method underestimates the amount of excess temperature in dry convection.*

(2) If both the ascending and the descending masses are saturated (e.g. within the cloud)

$$\Gamma = \Gamma' = \gamma_s$$

and the excess temperature of the ascending air is

$$\Delta T = \Delta z \left(1 + \frac{M'}{M}\right)(\gamma - \gamma_s)$$

$$= \Delta z \left(1 - \frac{w}{w'}\right)(\gamma - \gamma_s) \quad \ldots \quad (18)$$

In the parcel method $w = o$, and therefore

$$\Delta T = \Delta z\,(\gamma - \gamma_s) \quad \ldots \quad (18a)$$

It will thus be seen that *the parcel method and the slice method give the same sign of the excess temperature, but the parcel method underestimates the amount of excess temperature in saturated convection.*

(3) If the ascending air is saturated and the descending air non-saturated (as is the case when we consider a cumulus cloud relative to its environment), the excess temperature of the ascending air is

$$\Delta T = \Delta z \left[\gamma - \gamma_s - \frac{M'}{M}(\gamma_d - \gamma)\right]$$

$$= \Delta z \left[\gamma - \gamma_s + \frac{w}{w'}(\gamma_d - \gamma)\right] \quad \ldots \quad (19)$$

In the parcel method $w = o$ and, therefore,

$$\Delta T = \Delta z\,(\gamma - \gamma_s) \quad \ldots \quad (19a)$$

It will thus be seen that *the parcel method may, or may not, give the right sign of the excess temperature of the ascending air, and the parcel method greatly overestimates the amount of excess temperature.**

It will be seen from eq. (19) that the excess temperature vanishes and the stratification becomes neutral when

$$\frac{M'}{M} = \frac{\gamma - \gamma_s}{\gamma_d - \gamma} \quad \ldots \quad (20)$$

or

$$-\frac{w}{w'} = \frac{\gamma - \gamma_s}{\gamma_d - \gamma} \quad \ldots \quad (20a)$$

This shows that, for any given lapse rate, there is an upper limit of the ratio M′/M such that, if this ratio is exceeded, the stratification becomes stable even when $\gamma > \gamma_s$ in contradiction to the parcel method (i.e. eq. (19a)).

7. The Slice Method : The Positive Area

It is readily seen from Fig. 4 that the slice method, when applied to cloud-producing convective currents, renders positive areas that are much smaller than those indicated by the parcel method. According to the parcel method, the positive area of the parcel that has ascended the small distance from A to C, is represented by the triangle ACD, whereas the positive area according to the slice method, is represented by the triangle ACE. From this it follows that the vertical velocities and accelerations computed by the aid of the parcel method are much too large when saturated air ascends through a dry-adiabatically descending environment. It will also be seen that when the ratios represented by eqs. (20) are exceeded, there will be a negative area according to the slice method, although the parcel method may give a very large positive area.

It is of interest to compare the results of the slice method with those obtained by the parcel method as far as the classification of the ascent curves is concerned. We return to Fig. 2 (para. 3) and note that at the level D

$$\gamma = \gamma_s$$

If convection occurs below the level D, and a rising mass approaches this level, the air at the level D will be given an upward impulse, but according to eq. (19) this air will be stable and resist upward motion. In fact, stability would occur even below the level D, for the air would be stable for all impulses where

$$\frac{M'}{M} > \frac{\gamma - \gamma_s}{\gamma_d - \gamma}$$

Since in the layer just below D, γ differs only slightly from γ_s, this layer would be stable, except for rising masses which are exceedingly small as compared with the descending masses. Above the level D, the air would be stable for all impulses from rising masses from below, with the result that the part of the positive area above the level D in Fig. 2 does not become available for the rising masses from below.

With regard to the positive area shown in Fig. 2, the above discussion has shown :

(1) *that the positive area below the level* D *is greatly reduced as a result of the adiabatic heating of the environment,* and

(2) *that the positive area above the level* D *is fictitious inasmuch as it does not become available for the creation of convective currents.*

There is also a third factor in operation that tends to reduce the positive area of an ascending mass. A certain amount of mixing of heat and moisture will occur between the rising and sinking elements, with the result that the excess temperature and the corresponding positive area are reduced relative to the values indicated by the parcel method. The statistical results in Section III seem to indicate that less than 25 per cent. of the positive area shown in Fig. 2 actually becomes available for the creation of convective currents.

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* Except when the lapse rate within the convective cloud exceeds the dry-adiabatic lapse rate.

In view of the results obtained by the aid of the parcel method with regard to the top of the convective clouds and the penetration of the cloud tops into the upper negative area, as discussed in para. 4, it is of interest to examine in greater detail the processes above the level D in Fig. 2.

We consider a rising mass of saturated air in the layer below D. As this mass imparts an upward impulse to the air at the level D, the lapse rate at the level D may change. It is, therefore, of interest to determine whether the stability of the layer at, and above, D will increase or decrease with time.

In saturated-adiabatic motion the temperature change per unit time at a fixed point is

$$-\frac{\partial T}{\partial t} = w\,(\gamma_s - \gamma)$$

Differentiating with regard to z and noting that

$$-\frac{\partial^2 T}{\partial z \partial t} = \frac{\partial \gamma}{\partial t}$$

we obtain

$$\frac{\partial \gamma}{\partial t} = \frac{\partial w}{\partial z}\,(\gamma_s - \gamma) + w\,\frac{\partial}{\partial z}\,(\gamma_s - \gamma) \quad \ldots \quad \ldots \quad \ldots \quad (21)$$

But at the level D $\gamma = \gamma_s$, and therefore

$$\frac{\partial \gamma}{\partial t} = w\,\frac{\partial}{\partial z}\,(\gamma_s - \gamma) \quad \ldots \quad \ldots \quad \ldots \quad \ldots \quad (22)$$

Since $(\gamma_s - \gamma)$ increases from below upwards through the layer D,the lapse rate at D will increase when the saturated ascent reaches the level D, i.e. the stable layer will tend to ascend. However, on account of the stability above the level D, $\frac{\partial w}{\partial z}$ must be negative, and immediately above the level D the first term in eq. (21) comes into play, and its contribution is to reduce the lapse rate. At still higher levels $w = o$ and the first term in eq. (21) takes control, with the result that there is a tendency for the convective currents to create a stable layer, or an inversion, at some small distance above the layer D.

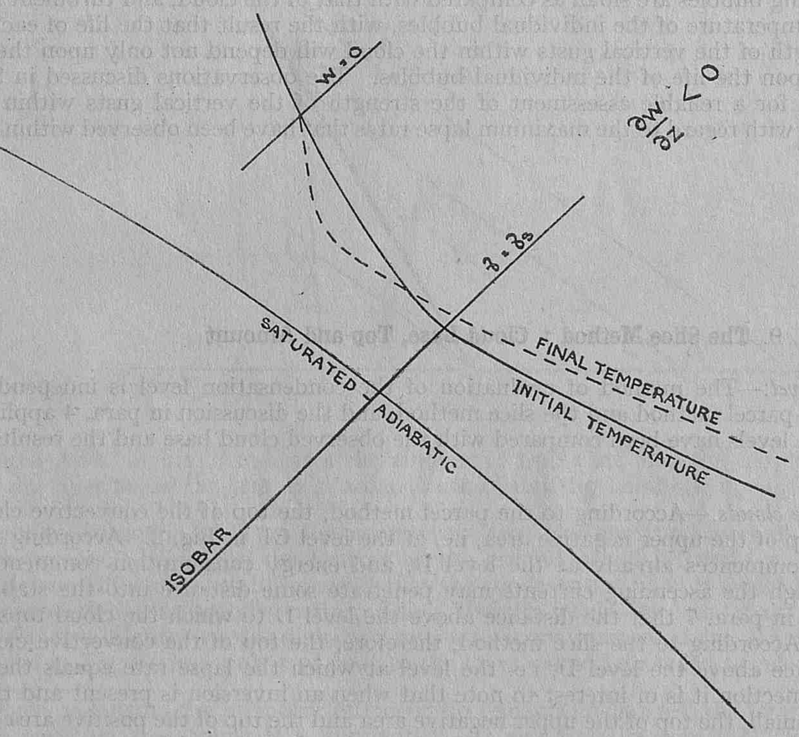

FIG. 5.—Schematical section of a TePhi-gram showing the change of stability in the upper part of the convective layer.

Thus, each convective current that approaches the level D in Fig. 2, will tend to lift this level slightly, and, at the same time, create increased stability in the layer above D. In Fig. 5 the full curve indicates the initial ascent curve, and the pecked curve the temperature distribution that results from repeated attacks by saturated convection from below.

8. **Further Remarks on the Vertical Velocity**

It was shown in the foregoing paragraphs that the parcel method underestimates the excess temperature and the vertical accelerations and velocities below the cloud base, and also within the cloud, whilst it greatly exaggerates the excess temperature and the vertical accelerations and velocities of the cloud air relative to the environment.

Experience shows that the convective cloud is not a uniformly ascending mass but a complex system of irregular currents. Although the mean velocity through a horizontal section of the cloud is upward, there are a number of downward currents alternating with upward gusts of considerable intensity. The life of each of the upward and downward moving elements within the cloud is short as compared with that of the cloud as a whole. Distinction must therefore be made between the general upward current of the cloud and the superimposed gusts.

The strength of the general upward current can be assessed roughly in two different ways.

(1) Take a cloud that has just formed. Even in pronounced cases it is rarely observed that the top of the growing cumulus rises more rapidly than 2 Km. per hour, or 0·6 m/sec., even when a large positive area is present. As will be shown in Section III, a velocity 0·6 m/sec. is only about 6 per cent. of the velocity that would be obtained by the parcel method, showing that the vertical velocities obtained by the parcel method for saturated ascent through a dry-adiabatically descending environment are much too large.

(2) Let I be the mean rain intensity per unit area under the cloud, a the mean absolute humidity below the cloud base, and let w be the mean upward velocity through the cloud base. When the steady state is reached, i.e. when the cloud neither grows nor accumulates water

$$\mathrm{I} = aw$$

If the rain intensity is 36 mm. per hour, and a is 15 gm/m^3

$$w = 0 \cdot 7 \text{ m/sec.}$$

which agrees with the value indicated above.

Since the above rain intensity is very large as a mean value under a whole cloud while the indicated absolute humidity is not excessive, it is reasonable to assume that the mean velocity of the cloud mass is not likely to exceed about 0·7 m/sec. To create such velocities, only a small difference between the virtual temperature of the cloud and the surrounding air is required.

With regard to the gusts that are often observed below the cloud base, and also within the cloud, different considerations hold, for the parcel method underestimates the amount of available energy when both the ascending and descending currents follow the same adiabatic.

Under the cloud base the lapse rate is normally only slightly super-adiabatic, with a result that the positive area below the cloud base is very small and the vertical gusts not excessive.

However, when the ascending air, with a slightly super-adiabatic lapse rate, reaches the condensation level, a relatively large positive area is immediately available, and large accelerations occur. On the other hand, the volumes of the rising bubbles are small as compared with that of the cloud, and turbulent mixing will rapidly reduce the excess temperature of the individual bubbles, with the result that the life of each bubble is relatively short. The strength of the vertical gusts within the cloud will depend not only upon the lapse rate within the cloud but also upon the life of the individual bubbles. The observations discussed in Section III are not sufficiently detailed for a reliable assessment of the strength of the vertical gusts within the cloud, but some statistics are given with regard to the maximum lapse rates that have been observed within convective clouds.

9. The Slice Method : Cloud Base, Top and Amount

(1) *The condensation level.*—The method of evaluation of the condensation level is independent of the assumptions underlying the parcel method and the slice method, and the discussion in para. 4 applies to both. The evaluated condensation levels have been compared with the observed cloud base and the results are given in Section III.

(2) *The top of convective clouds.*—According to the parcel method, the top of the convective cloud should be found at, or near, the top of the upper negative area, i.e. at the level GF in Fig. 2. According to the slice method, the stable layer commences already at the level D, and energy consumption commences already below the level D. Although the ascending currents may penetrate some distance into the stable layer, it follows from the discussion in para. 7 that the distance above the level D to which the cloud tops can reach must be relatively small. According to the slice method, therefore, the top of the convective cloud will be found at some small distance above the level D, i.e. the level at which the lapse rate equals the saturated adiabatic rate. In this connection it is of interest to note that when an inversion is present and the positive area below the inversion is small, the top of the upper negative area and the top of the positive area will not be very far above the level D. In these circumstances the slice method and the parcel method will render results that do not differ greatly. As will be shown in Section III, very large differences are found when the stability above the level D is slight.

(3) *Cloud amount.*—It was shown in para. 6 that, for any given lapse rate, there is a value of the ratio of the ascending mass to the descending mass which, when exceeded, causes the stratification to become stable even when the lapse rate exceeds the super-adiabatic rate. This ratio is expressed by eq. (20). Since M′ and M denote the masses contained within a slice of unit thickness, the ratio of the masses may be replaced by the ratio of the areas involved. Accordingly, the limit at which energy production ceases is expressed by

$$\frac{\mathrm{A}'}{\mathrm{A}} = \frac{\gamma - \gamma_s}{\gamma_d - \gamma}$$

where A′ is the horizontal area through which the air ascends, and A is the horizontal area through which the air descends. This ratio applies to the energy-producing clouds, i.e. the clouds that have a temperature in excess of that of the environment. In addition to the energy-producing clouds, passive, or energy-consuming clouds may be present. However, if convection is active, the energy-consuming and the passive clouds cannot exist in great amounts, for these clouds will dissolve in the descending currents. As a result of this, the amount of clouds in active convection is not likely ot exceed much the amount indicated by the above ratio.

In this connection it is of interest to note that the cloud amount assessed by the ground observer represents a projection of the cloudy areas on the sky. These cloud amounts are therefore exaggerated on account of the perspective effect. The air observer, on the other hand, will report cloud amounts representing more nearly the true ratio of the cloudy and the clear spaces. Thus, the amounts reported by the air observer are largely comparable with the ratio referred to above.

In typical situations the lapse rate in the convective cloud layer is rarely greater than

$$\tfrac{1}{2}\,(\gamma_d + \gamma_s)$$

with a result that the ratio of the cloudy to the clear area is not likely to exceed unity, and the observed cloud amounts (assessed without the perspective effect) are not likely to exceed five-tenths. This holds for active convection (i.e. energy-producing clouds). When an inversion is present, the convective clouds spread out under it in the energy-consuming layer. The typical cloud form is then strato-cumulus, or a layer of flat cumulus, and the amount of these clouds often exceeds five-tenths of the sky.

10. **The Convection Temperature**

We consider a stable layer of air above the ground as shown in Fig. 6. If the air at A were lifted adiabatically and with constant moisture content, it would become saturated at its condensation level C. If the air at A is heated with constant moisture content, the level C will rise, and by the time the surface temperature has reached the value TC, the condensation level will have ascended to the point B on the ascent curve. If, at this time, a dry-adiabatic lapse rate has been established below the level B, convection from the ground would be released and the ascending air would reach the condensation level B. The temperature indicated by TC in Fig. 6 is called the *convection temperature*; its evaluation from the TePhi-gram is evident from Fig. 6.

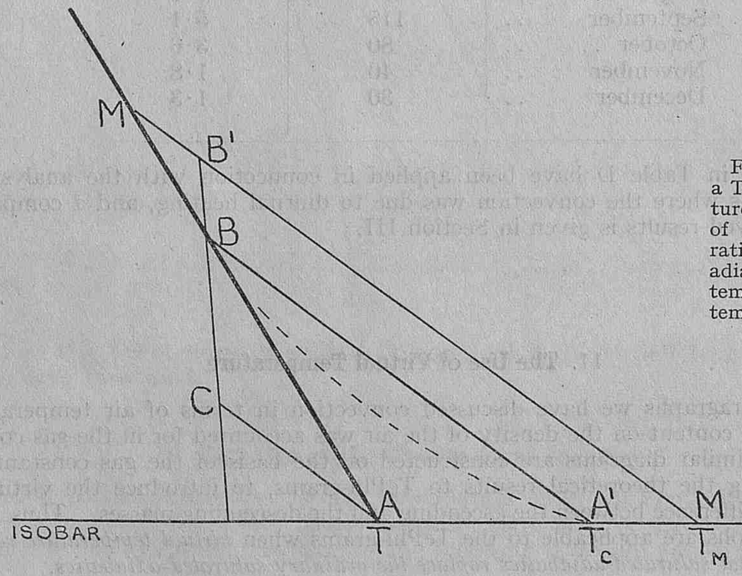

FIG. 6.—Schematical section of a TePhi-gram. AB = temperature curve at sunrise; CB = line of constant humidity mixing ratio; AC, A′B and MM = dry-adiabatics; T_C = convection temperature; T_M = maximum temperature.

The area ABA′ in Fig. 6 indicates the amount of heat that must be supplied per column of unit cross section of the layer below the level B in order to establish a dry-adiabatic lapse rate down to the ground.

The above deduction rests on the assumption that the moisture content of the air remains constant during the process. Normally, the heating of the ground would intensify evaporation, and the humidity mixing ratio would increase, with the result that the condensation level B would be found at some lower level, and the convection temperature corresponding thereto would be less than that indicated in Fig. 6. Furthermore, if a slightly super-adiabatic lapse rate is maintained above the ground (as indicated by the pecked curve in Fig. 6), less supply of heat to the layer would be needed to release convection.

However, as long as the humidity mixing ratio remains unchanged, the temperature TC must be reached before the negative area below B is eliminated. It will thus be seen that the area ABA′ represents the maximum amount of energy that must be supplied prior to the appearance of convective clouds. A super-adiabatic lapse rate above the ground will substantially reduce this area, but the temperature TC would not be affected as long as the humidity mixing ratio does not change.

In the early morning there is often a minimum of moisture near the ground. In such cases a representative value of the humidity mixing ratio must be used. It has been customary to assume that the humidity mixing ratio will mix and become uniform along the vertical by the time convection commences, with the result that the mean humidity of the layer rather than the observed values at the ground should be used for assessing the convection temperature.

The convection temperature is of considerable use in the forecasting of convection due to diurnal heating, and the success that can be obtained depends largely upon the ability of the forecaster to estimate whether or not the actual mid-day temperature will reach (or exceed) the critical value. In this assessment the forecaster will be guided not so much by the difference TC − T in Fig. 6 as by the size of the area ABA′, for if the lapse rate is steep, the heating will be spread over a deep layer, and the diurnal amplitude of temperature will be small.

An assessment of the amounts of heat supplied to the surface layer through diurnal insolation has been made by Gold (1933), who converted the amounts of heat into areas on the TePhi-gram. This conversion is not quite exact, but it is sufficiently accurate in comparison with the errors introduced by the observational material and the assumptions underlying the derivation. The values obtained by Gold are given in Table D. As regards the method of derivation, etc., the reader is referred to the original paper.

The areas in Table D represent the amounts of energy supplied during the period from sunrise until the maximum temperature is reached. This area is represented by AMM in Fig. 6, where T_M indicates the diurnal maximum temperature. It will be seen that if the lapse rate above the ground is super-adiabatic, a substantially smaller area will give the same maximum temperature. In this connection it may be mentioned that Swinbank (1944), on the basis of more plausible assumptions as to the exchange of heat between the air and the ground, finds amounts of energy which are smaller than those given in Table D. However, when the energy values found by Gold and Swinbank are converted into maximum temperature, the results are in good agreement.

TABLE D

Monthly Values of Available Heat from Sunrise to early Afternoon and the corresponding Areas on the TePhi-gram (according to Gold)

Month	gm. cal/cm²	Area (cm²)
January	40	1·8
February ..	70	3·1
March	100	4·4
April	140	6·2
May	175	7·8
June	180	8·0
July	165	7·3
August	150	6·7
September ..	115	5·1
October	80	3·6
November ..	40	1·8
December ..	30	1·3

The areas indicated in Table D have been applied in connection with the analysis of a number of TePhi-grams in situations where the convection was due to diurnal heating, and a comparison between the computed and the observed results is given in Section III.

11. The Use of Virtual Temperature

In the foregoing paragraphs we have discussed convection in terms of air temperature, i.e. as if the influence of the moisture content on the density of the air was accounted for in the gas constant. Since the TePhi-gram and other similar diagrams are constructed on the basis of the gas constant for dry air, it is necessary, when applying the theoretical results to TePhi-grams, to introduce the virtual temperature to account for the density difference between the ascending and the descending masses. Thus, the results derived in the foregoing paragraphs are applicable to the TePhi-grams when *virtual temperature replaces the dry-bulb temperature, and the virtual saturated-adiabatics replace the ordinary saturated-adiabatics.*

(1) *The virtual temperature.*—The total density of the moist air is the sum of the density of the dry air and of the aqueous vapour. If ϱ is the density of the moist air, p the atmospheric pressure, and e the vapour pressure

$$\varrho = \frac{p - e}{RT} + 0{\cdot}622\ \frac{e}{RT} = \frac{p}{RT}\left(1 - 0{\cdot}378\ \frac{e}{p}\right)$$

where R is the gas constant of dry air, and T the Absolute temperature. The virtual temperature T_v is then defined as the temperature that the dry air will have if its density were ϱ and its pressure p. Hence

$$T_v = \frac{T}{1 - 0{\cdot}378\ \dfrac{e}{p}}$$

Now, the humidity mixing ratio is

$$q = \frac{0{\cdot}622\ e}{p - e}$$

Since q is of the order of 10^{-2}, second and higher powers of q may be neglected. With sufficient accuracy therefore

$$T_v = T\,(1 + 0{\cdot}0006\ q)$$

where T_v and T are expressed in °K and q is expressed in grammes per kilogramme of dry air.

When the air temperature is below 0° F. the humidity mixing ratio is so small that the virtual temperature increment falls below the observational accuracy. At these low temperatures, therefore, the virtual temperature increment may be neglected. At high temperatures, however, the virtual temperature increment becomes important. For example, at 90° F. and 1000 mb. the virtual temperature increment may be as large as 10° F. If the air outside the cloud is relatively dry, large density contrasts may exist on account of the difference in the humidity inside and outside the cloud. In such cases the cloud air may be lighter than the environment although the air within the cloud may be colder than the surrounding air.

(2) *The virtual saturated-adabiatics.*—We consider a schematical TePhi-gram as shown in Fig. 7. A parcel of air has the temperature T, the wet-bulb temperature T_w, and the virtual temperature T_v. If this parcel ascends, T and T_v will vary dry-adiabatically until the condensation level CL is reached. It will be seen that the condensation level is evaluated from the dry-bulb and the wet-bulb temperatures, and not from the virtual temperature.

As the parcel ascends above the condensation level, the temperature point will follow the saturated-adiabatic S_2, but the virtual temperature will *not* follow the saturated-adiabatic S_1, for, as water is condensed, the virtual temperature increment will decrease and approach zero at temperatures below about 0° F. Hence, the virtual temperature of the ascending saturated air will decrease as indicated by the curve S_v in Fig. 7, i.e. the *virtual saturated-adiabatic*.

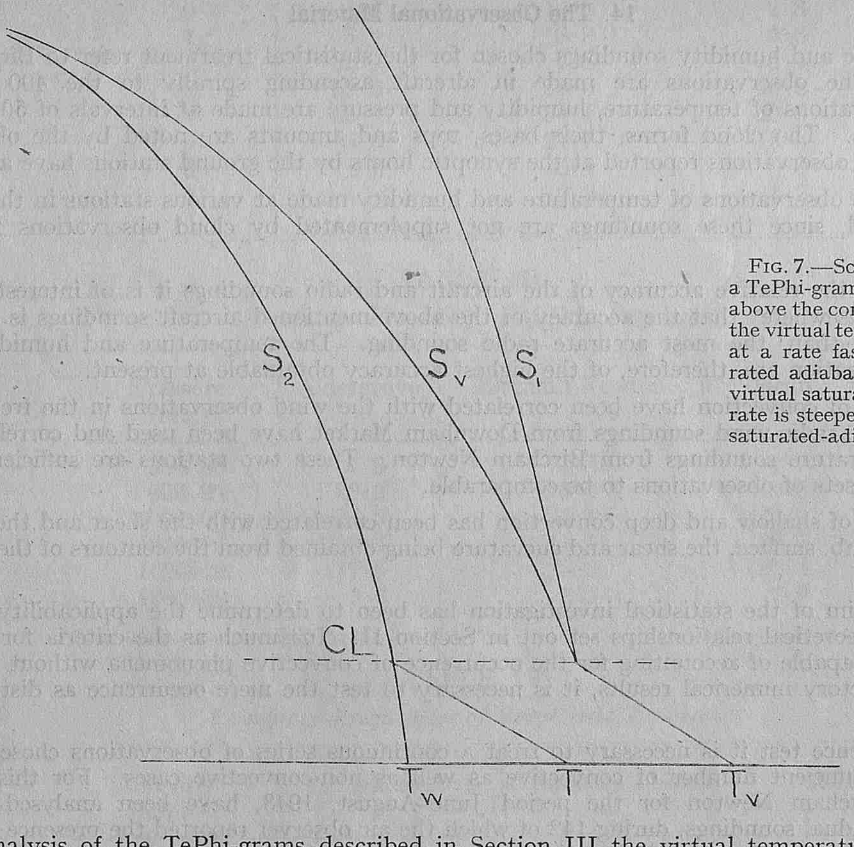

FIG. 7.—Schematical section of a TePhi-gram. As the air ascends above the condensation level CL, the virtual temperature decreases at a rate faster than the saturated adiabatic S_1. Hence, the virtual saturated-adiabatic lapse rate is steeper than the ordinary saturated-adiabatic rate.

In the analysis of the TePhi-grams described in Section III the virtual temperature and the virtual saturated adiabatics have been used.

12. **Remarks on the Observational Accuracy**

For the evaluation of such quantities as the condensation level, the convection temperature and the positive and the negative areas, the surface observations are used, and the errors in these observations will influence the accuracy of the evaluated quantities. The following is readily obtained from a TePhi-gram.

To an error of ± 1° F. in the surface dry-bulb temperature corresponds : (1) an error of ± 1° F. in the virtual temperature, (2) no error in the convection temperature, and (3) an error of ∓ 10 to 15 mb. in the condensation level. In convective situations the negative area is either very small or absent. An error of ± 1° F. in the surface dry-bulb temperature will then have no appreciable influence on the accuracy of the evaluated positive area, and the error in the negative area is proportional to the distance to the condensation level.

To an error of ± 1° F. in the surface wet-bulb temperature corresponds ; (1) No appreciable error in the virtual temperature, (2) a ∓ error in the convection temperature, the magnitude of which varies greatly with the lapse rate, (3) an error of ± 10 to 15 mb. in the condensation level, (4) a ∓ error in the negative area, the magnitude of which varies with the ascent curve, and (5) a ± error in the positive area, the magnitude of which is proportional to the length of the positive area.

It is readily seen (e.g. from Fig. 2) that as far as the positive area is concerned, an error of ± 1° F. in the surface wet-bulb temperature has the same effect as an error of ± 1° F. in the dry-bulb temperature throughout the air column. It will thus be seen that a high degree of accuracy is required in the surface humidity observations.

13. **Release of Precipitation**

According to Bergeron (1933), the presence of ice crystals and super-cooled water droplets in the clouds is an essential condition for colloidal instability. The instability is due to the depression of the saturation vapour pressure over ice, which has a maximum at about 10° F. When the temperature in the upper part of the cloud is appreciably below freezing, the cloud would become colloidably unstable and precipitation would be released.

Computations show (Petterssen 1940) that the growth of the cloud droplets through condensation decreases rapidly when the drops increase, and the colloidal instability due to ice crystals can only account for the initial growth from ice crystals to large cloud elements, but is insufficient to account for the size of the raindrops. However, as the larger cloud elements will fall more rapidly than the small elements, coalescence through collisions will occur, and the falling drops will increase in proportion to the depth of the cloud through which they fall.

A number of ascents made in convective situations with precipitation have been analysed and classified according to the temperature in, and the depth of, the clouds. The results are given in Section III.

III. STATISTICAL CHARACTERISTICS

As far as the observational material has been found suitable, the theoretical results referred to in the foregoing section have been tested against the analysis of synoptic charts and the observations of temperature, humidity, etc., in the free atmosphere. The results of this comparison are summarised in the following paragraphs.

14. The Observational Material

The temperature and humidity soundings chosen for the statistical treatment refer to Bircham Newton and Aldergrove. The observations are made in aircraft ascending spirally to the 400 mb. surface. Simultaneous observations of temperature, humidity and pressure are made at intervals of 50 mb. and also at significant points. The cloud forms, their bases, tops and amounts are noted by the observer in the aircraft. The cloud observations reported at the synoptic hours by the ground stations have also been used.

The radio-sonde observations of temperature and humidity made at various stations in the British Isles have not been used, since these soundings are not supplemented by cloud observations from the free atmosphere.

With regard to the relative accuracy of the aircraft and radio soundings it is of interest to note that it has been shown elsewhere* that the accuracy of the above-mentioned aircraft soundings is as high as, or even slightly higher than, the most accurate radio sounding. The temperature and humidity soundings used for this investigation are, therefore, of the highest accuracy obtainable at present.

Certain aspects of convection have been correlated with the wind observations in the free atmosphere. For this purpose the radio-wind soundings from Downham Market have been used and correlated with the simultaneous temperature soundings from Bircham Newton. These two stations are sufficiently near one another for the two sets of observations to be comparable.

The occurrence of shallow and deep convection has been correlated with the shear and the curvature of the flow in the 750 mb. surface, the shear and curvature being obtained from the contours of the said pressure surface.

The principal aim of the statistical investigation has been to determine the applicability to upper air soundings of the theoretical relationships set out in Section II. Inasmuch as the criteria for stability and instability may be capable of accounting for the occurrence of convective phenomena without being capable of rendering satisfactory numerical results, it is necessary to test the mere occurrence as distinct from the numerical fit.

For the occurrence test it is necessary to treat a continuous series of observations chosen at random, which contains a sufficient number of convective as well as non-convective cases. For this purpose the soundings from Bircham Newton for the period June–August, 1943, have been analysed. This series comprises 276 individual soundings, during 142 of which the air observer reported the presence of convective clouds. This series of observations will be referred to as the *non-selected series.*

For the test of the numerical accuracy of the various criteria and theoretical relationships the soundings from Bircham Newton and Aldergrove for the period April–November, 1943, have been used, and 295 cases of convection were chosen for investigation. The selection of convective situations was made by inspection of the sea level synoptic charts, and a situation was selected for investigation, if it was found to satisfy the following conditions :—

(1) Convective clouds with or without precipitation reported at or near Bircham Newton or Aldergrove.
(2) No front present within 200 miles of these stations.

This selection aims at the exclusion of convection released by fronts, i.e. convection that does not necessarily commence at the ground. A detailed analysis of the individual soundings showed that a few of the selected cases represented convection due to fronts which were so feeble that they had not been introduced in the surface analysis. These cases will be discussed separately.

It is of interest to note that the soundings themselves were not used for selecting the convective situations, nor was any attempt made at selecting according to intensity of convection, type of synoptic situation, etc. The statistical results should therefore, represent a fair cross-section of the conditions typical for convection within the British Isles. Finally, it may be noted that mid-winter situations were excluded on account of the reduced quality of the cloud observations at the times of the morning and evening ascents.

15. Characteristics of the Selected Situations

Certain statistical characteristics of the selected synoptic situations at sea level are given in Tables I–V.

TABLE I

Percentage Frequency of Wind Directions at Station Level

	Aldergrove	Bircham Newton	Mean
NE	0·6	1·6	1·1
E	1·8	0·8	1·3
SE	3·6	0·8	2·2
S	4·8	2·3	3·5
SW	13·2	14·7	14·0
W	30·2	30·2	30·2
NW	32·0	27·1	29·5
N	9·0	21·7	15·4
Calm ..	4·8	0·8	2·8

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* M.R.P. 204.

TABLE II

Percentage Frequency of Wind Force at Station Level

Force	Aldergrove	Bircham Newton	Mean
0	4·8	0·8	2·8
1	5·4	3·9	4·6
2	22·9	7·8	15·4
3	29·6	24·0	26·8
4	26·5	34·1	30·3
5	9·6	21·7	15·6
6	1·2	5·4	3·3
7	—	2·3	1·2

TABLE III

Percentage Frequency of M.S.L. Pressure

Pressure (mb.)	Aldergrove	Bircham Newton	Mean
980–89	1·8	1·5	1·7
990–99	13·2	5·4	9·3
1000–09	23·5	26·4	25·0
1010–19	40·4	41·1	40·7
1020–29	17·5	23·3	20·4
1030–39	3·6	2·3	2·9

TABLE IV

Percentage Frequencies of Barometric Tendencies

Tendency (mb.)	Aldergrove	Bircham Newton	Mean
⩾3·0	7·8	3·2	5·5
2·0 to 2·9	4·2	10·0	7·1
1·0 to 1·9	24·7	17·8	21·3
0·0 to 0·9	32·1	32·5	32·3
−0·1 to −1·0	19·8	27·8	23·8
−1·1 to −2·0	9·6	4·7	7·1
−2·1 to −3·0	1·8	1·6	1·7
<−3·0	—	2·4	1·2
Mean Tendency ..	0·6	0·5	0·6

TABLE V

Percentage Frequency of Curvature of M.S.L. Isobars

Curvature	Aldergrove	Bircham Newton	Mean
Cyclonic	36	42	39
Straight	33	37	35
Anticyclonic	31	21	26

It will be seen from Table I that about 75 per cent. of the selected cases occurred with surface wind directions from W, NW or N, suggesting a preponderance of polar outbreaks. The relatively high wind forces shown in Table II suggest that advection rather than diurnal heating is, in the majority of cases, responsible for the creation of instability, although diurnal heating contributes to the creation of steep lapse rates. The times to which the selected ascents refer are shown in Table VI.

TABLE VI

Time of Observation	Percentage Frequency of Ascents
Morning	20
Midday	45
Evening	35

It will be seen from Table VI that at least 60 per cent. of the selected cases represent convection due to advective heating. The remaining 40 per cent. will be due to either pure diurnal heating or diurnal heating assisted by advection.

It will be seen from Table IV that about two-thirds of the selected cases occurred together with rising pressure, and this together with Table I suggests that the majority of the selected cases belong to the rears of travelling depressions, or polar currents spreading eastward. Table V indicates that, on the whole, there is a slight preference for cyclonically curved isobars. In this connection it is possibly significant that the preference for cyclonically curved sea level isobars is higher at Bircham Newton, where the length of the land track is great, than it is at Aldergrove where the length of the land track is very short.

It is important to bear in mind that the characteristics shown in the above tables do not result from the method of selection of cases, for the cases were selected on the criteria mentioned in para. 14, without regard to the synoptic situation, type of circulation, etc. The above characteristics may therefore be taken to indicate the conditions that normally are favourable for the development of convective phenomena over the British Isles.

16. **Remarks on Cloud Observations**

The cloud forms and weather phenomena which are generally recognised to result from convective currents are :—

Cumulus humilis (Cu)
Cumulus congestus (CuC)
Cumulo-nimbus (Cb)
Showers, squalls, thunderstorms and hail.

In many cases it has been found that convection results in strato-cumulus (Sc), particularly when the convective layer is limited upward by a stable layer, or an inversion. Thus, strato-cumulus together with cumulus will often be found to occur. Quite frequently the convective clouds may appear as well-defined cumuli as seen from the ground, but as a layer of strato-cumulus as seen from the aircraft. This is particularly true when the tops of the individual cumuli appear at a uniform level. In these cases, the ground observer and the air observer may report different types of clouds.

While the ground observer makes distinction between cumulus humilis (i.e. flat cumulus) and cumulus congestus (i.e. towering cumulus), no such distinction is made by the air observer who reports the height of the base and of the top of the cumuli. In the following, no distinction will be made between cumulus humulis and cumulus congestus, both being denoted by the symbol Cu.

The distinction between cumulus congestus and cumulo-nimbus is also difficult, for the tops of these clouds cannot always be seen by the ground observer who, in his assessment, is guided by the appearance of the sky as a whole rather than by the structure of the individual clouds.

A comparison has been made between the cloud forms reported by the ground observers and the air observers in the 295 cases selected for investigation, and the results are shown in Table VII.

TABLE VII

Percentage Frequency of Cloud Forms reported by the Ground Observer and the Air Observer. Selected Series.

Air Observer	Ground Observer				Total
	Cu	Cb	Cu and Sc	Sc	
Cu	20	6	7	2	35
Cb	4	18	1	—	23
Cu and Sc	9	2	8	2	21
Sc	8	1	6	6	21
Total	41	27	22	10	100

It will be seen from Table VII that complete agreement between the two observers is obtained only in 52 per cent. of the cases. However, an examination of the individual cases shows that the discrepancies represent borderline cases in which it has been difficult to distinguish between flat cumulus and strato-cumulus, or between towering cumulus and cumulo-nimbus, or between cumulus with and without strato-cumulus. Except in three cases (i.e. 1 per cent.) where the air observer reported Sc while the ground observer reported Cb, the cloud reports by both observers have been found satisfactory.

In 6 per cent. of the cases both the air observer and the ground observer reported Sc without cumulus or other instability phenomena. In these cases it was found that neighbouring stations and the synoptic charts indicated a convective situation. These cases have therefore been included ; they may be taken to represent convection within a shallow layer.

17. **Occurrence of Convective Clouds**

In order to determine the extent to which the parcel method and Normand's classification of ascent curves (as referred to in para. 3) are capable of accounting for the occurrence of convective clouds, the non-selected series of observations referred to in para. 14 has been used, and the results of the analysis of the ascent curves have been compared with the cloud observations made from the aircraft. In the analysis the negative and the positive areas were evaluated by the use of virtual temperatures and the virtual saturated-adiabatics, as explained in para. 11.

The non-selected series covers the period June–August, 1943. It refers to Bircham Newton and contains 270 ascents with cloud observations.

If, following Normand, the criterion for instability is that the positive area is greater than the negative area, and, similarly, the criterion for stability is that the negative area is greater than the positive area, the 270 ascents with cloud observations can then be referred to the following four categories, viz. :—

A. The positive area greater than the negative area with clouds of the cumulus type.*
B. The positive area smaller than the negative area without clouds of the cumulus type.
C. The positive area greater than the negative area without clouds of the cumulus type.
D. The positive area smaller than the negative area with clouds of the cumulus type.

The percentage frequency of the above categories were found to be as shown in Table VIII.

TABLE VIII

Category	Percentage Frequency
A	43
B	25
C	21
D	11
A + B	68
C + D	32

It will be seen from Table VIII that the cloud observations confirmed the criterion for instability in 43 per cent. of the cases (i.e. category A), and confirmed the criterion for stability in 25 per cent. of the cases. Thus, in all, the criteria gave satisfactory results in 68 per cent. of all cases. The frequencies in the categories C and D represent the number of cases in which the cloud observations failed to confirm the criteria.

The failure of the criteria to render the expected results is not symmetrical, for the frequency in category C is about twice as large as that in category D, i.e. there is a small number of cases with cumulus clouds in contradiction to the criterion for stability, but a fairly large number of cases where the criterion for instability was satisfied without clouds of the cumulus type resulting.

Category B is of no further interest, for neither the criteria nor the observations indicate that any form of instability was present.

Category C is of considerable interest ; the criteria indicate that instability is present, but the observations show no cloud of the recognised convective type (i.e. Cu or Cb).

Of the 58 observations in category C, 48 showed strato-cumulus and 6 stratus ; in 4 cases no low clouds were observed.

We shall first consider the four cases in which the instability criterion was satisfied without low cloud being present. In three of these cases there was no negative area at all, i.e. super-adiabatic lapse rate above the ground. However, the air was so dry that the condensation level was well above the top of the positive area, with the result that no cloud could form. In the fourth case the positive area was slightly larger than the negative area, but there was a strong inversion at the ground which, apparently, prevented convection from being released. The ascent referred to was made on 24th August, 1943, at 0600 G.M.T. At 1200 G.M.T. the diurnal heating had eliminated the negative area, and cumulus cloud had formed. Thus, on closer examination, the lapse rate criteria of the parcel method and of the slice method (see paras. 1 and 6) were fully satisfied.

We shall next examine the 54 cases in which strato-cumulus (48) and stratus (6) were reported in category C. The frequency of occurrence of these clouds in relation to the negative and positive areas is shown in Table IX.

TABLE IX

Frequency of Strato-cumulus and Stratus in Combinations of Positive and Negative Areas. Non-selected Series, Category C. (Figures in Parentheses refer to Stratus).

Positive area (cm²)	Negative Area (cm²)				Total
	0·0	0·1/0·2	0·3/0·4	0·5/0·6	
0·0	—				—
0·1–0·5	13 (2)	1	1		15 (2)
0·6–1·0	12 (1)		1	2	15 (1)
1·1–2·0	8			(1)	8 (1)
2·1–3·0	— (1)				— (1)
3·1–4·0	3				3
4·1–5	—				—
6–10	3 (1)				3 (1)
11–20	1				2
>20	1	1		1	2
Total	41 (5)	2	2	3 (1)	48 (6)

It will be seen from Table IX that 41 of the 48 cases of strato-cumulus and 5 of the 6 cases of stratus occurred without a negative area being present. In these cases, the lapse rate near the ground was dry-adiabatic or steeper, and there can be little doubt that the clouds were formed through convection. In the remaining cases (i.e. 8 out of 54) the negative area was so slight that, unless the temperature was extremely uniform horizontally, the negative area would be certain to disappear locally. Furthermore, even minor inaccuracies in the observations would give negative areas of the magnitude indicated in Table IX.

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* By clouds of cumulus type is here understood Cu and Cb, and not Sc.

TABLE X

Frequency of Clouds of the Cumulus Type in Combinations of Positive and Negative Areas.

Non-selected Series. Categories A and D.

Positive area (cm²)	Negative area (cm²)												Total	
	0	0·1/0·2	0·3/0·4	0·5/0·6	0·7/1·0	1·1/2·0	2·1/3·0	3·1/5	6/10	11/20	21/30	>30	A	D
0									1	4	1	8	0	14
0·1/0·2	5				1		2		1		1	2	5	7
0·3/0·4	3	1				1	1			1			4	3
0·5/0·6	4					1	1	1		1			4	4
0·7/1·0	5	2	3	2								1	12	1
1·1/2·0	12	2	3		2								19	
2·1/3·0	11	2											13	
3·1/5	11	1	1		2								15	
6/10	13	3	2	1									19	
11/20	13		2		3	1							19	
21/30	4												4	
>30	5		1										6	
Total A	86	11	12	3	7	1							120	
Total D					1	2	4	1	2	6	2	11		29

Note.—The upper right half of the table refers to Category D, and the lower left to Category A.

It is of interest to note that the positive areas in Table IX are relatively small, and this suggests that the unstable layers were relatively shallow, i.e. they were limited by stable layers, or inversions, at some distance above the ground.

From the above discussion it appears that the 54 cases represent convective strato-cumulus of the type referred to in para. 8. Thus, recognising this type of strato-cumulus as a convective cloud, it will be seen that the instability criteria of the parcel method and the slice method, as applied to the conditions between the earth's surface and the cloud base, account for the presence, or absence, of convective clouds in 89 per cent. of the cases referred to in Table VIII, i.e. the sum of the frequencies in the categories A, B and C.

Category D.—We shall next discuss the clouds in category D in greater detail and compare their characteristics with those of the clouds in category A. The clouds in these categories have been grouped according to the size of the positive and negative areas, and the results are shown in Table X.

It will be seen that the clouds of the category A are characterised by the absence of, or a very small, negative area and, on the whole, a very large positive area, which is in agreement with what one would expect. On the other hand, the clouds in category D are characterised by a very large negative area and a relatively small positive area. It is indeed difficult to see how convection could be released from the ground in the numerous cases where there is no positive area but a very large negative area.

In order to throw further light on this question we shall examine the amount and the depth of the clouds in categories A and D. If N denotes the number of tenths of the sky covered by cumulus cloud, and Δp the pressure difference between the base and the top of the cloud, we shall define the *equivalent cloud amount* E by

$$E = \frac{N \Delta p}{50}$$

The figure 50 in this formula is introduced as a convenient unit for measuring the depth of the cloud. In the above formula N is the cloud cover as reported by the air observer which, unlike the cloud cover reported by the ground observer, is largely free from the perspective effect. The equivalent cloud amount is readily interpreted. Thus, $E = 2$ means that if the clouds were arranged in a column 50 mb. deep, two tenths of the sky would be covered, and similarly, if $E = 20$, there would be sufficient cloud to fill a 100 mb. layer with complete overcast, or a 200 mb. layer with half overcast.

The equivalent amounts of cumulus clouds belonging to the categories A and D have been correlated with the difference ΔA between the positive and negative areas as shown in Table XI, where $\Delta A > 0$ indicates that the positive area is greater than the negative area (i.e. category A), and $\Delta A < 0$ indicates that the negative area is greater than the positive area (i.e. category D).

TABLE XI

Frequency of Equivalent Amounts of Cumulus Cloud for Various Values of the Difference between Positive and Negative Areas. Non-selected Series. Categories A and D.

ΔA (cm^2)	Equivalent Cloud Amount										Total
	0·2	0·4	0·6	0·8	1·0	1·1/2·0	2·1/5·0	5·1/10·0	10·1/20·0	20	
> 30						1	1	1		3	6
21/30							1	2	1		4
11/20					1		2	2	4	6	15
6/10			2		1		1	6	4	4	18
3·1/5				1	1	7	2			1	12
2·1/3·0	1	2	1			2	3	4			13
1·1/2·0		1		1		2	5	3		1	13
0·7/1·0	2		2			1	3				8
0·5/0·6	1		2			2	1				6
0·3/0·4	1					1	3				5
0·1/0·2	1				3		2				6
0											
−0·1/−0·2											
−0·3/−0·4											
−0·5/−0·6	1										1
−0·7/−1·0		1					1				2
−1·1/−2·0	1	1			1						3
−2·1/−3·0			1							1	2
−3·1/−5						1					1
−6/−10		1									1
−11/−20	2	1				1		1			5
−21/−30						1		2			3
<−30	2	2	1	1		2		2			10
Total A	6	3	7	2	6	16	24	18	9	15	106
D	6	6	2	1	1	5	1	5	—	1	28

Note.—The upper half of the table refers to Category A and the lower half to Category D. This table contains 15 cases less than does Table X, the difference being due to missing observations of cloud top or cloud base.

It will be seen from Table XI that there is some positive correlation between ΔA and the equivalent cloud amount in category A, but no such correlation in category D. It will also be seen that the majority of the cases in category D represent insignificant equivalent amounts. Only in 7 of the recorded cases in category D are the equivalent amounts appreciable, and all these cases occurred with excessive stability near the ground. Although the possibility of erroneous cloud reports cannot be excluded, the number of such reports is not likely to account for the cases in category D, which represent 11 per cent. of all cases examined. An examination of the synoptic charts showed that most of the cases with high equivalent cloud amounts in category D were associated with fronts, and in these cases convection is not released from the ground, which explains the occurrence of heavy convective clouds although the negative area of the surface air was very large.

A further source of the cases in category D may be sought in the circumstance that in many cases, notably with small equivalent amounts, convection occurred at sea without the ascent at Bircham Newton showing instability in the surface layer. In this connection it is significant that of the 28 ascents in category D, 17 were made at about 0600, 2 at about 1200, and 9 at 1800 G.M.T.

Summary.—From the above discussion it follows that, with the exception of a few doubtful cases, the criteria of the parcel method and of the slice method for convection to be released from the ground have proved to yield results that are entirely satisfactory. Thus, the criteria are capable of accounting for the occurrence of the convective clouds that are due to convection from the ground. From this, however, it does not follow that the theories underlying the parcel method and the slice method are capable of accounting for other aspects of convection, such as the vertical velocities, the vertical and horizontal extent of convective clouds, etc. These questions will be examined by the aid of the selected cases to be analysed in the following paragraphs.

It will be seen from the Tables IX and X that of the 174 cases in the categories A and C (which certainly represent convection from the ground), about 75 per cent. occurred without a negative area near the ground, and in about 90 per cent. of the cases were the negative areas less than $0\cdot5$ cm^2. This seems to suggest that the condition for convection to be released can be expressed by the smallness of the negative area rather than by the difference between the size of the positive and the negative area (see also paras. 2 and 21).

18. Amount of Convective Clouds

It was mentioned in para. 4 that the parcel method does not give any information as to the amount of convective clouds that may develop for any given stratification. On the other hand, it was shown in para. 9 that according to the slice method there is, for any given lapse rate, an upper limit to the amount of *energy-producing clouds.* In addition to this amount, energy-consuming clouds may exist, but as long as convection is active, the descending currents around the energy-producing clouds will tend to eliminate the energy-consuming clouds, with the result that in the active phase of the convective process most of the clouds are energy-producing, whereas in the last phase, when the convection is dying out, large masses of passive, or energy-consuming clouds may be present in the sky.

Furthermore, it follows from the slice method that the upper limit for the amount of energy-producing cloud is about 5 tenths, except in rare cases when the lapse rate is excessively steep. On the basis of this, one would expect to find that, in the large majority of cases, the amount of cloud is 5 tenths or less.

The percentage frequency of the amounts of cloud of the cumulus type (i.e. Cu and Cb) as reported by the air observers at Bircham Newton and Aldergrove are given in Table XII, for the selected and the non-selected series of observations.

TABLE XII

Percentage Frequency of Amounts of Cloud of the Cumulus Type. Bircham Newton and Aldergrove.

Amount	Non-selected series	Selected series	Mean
Trace or 1/10	16	6	11
1/4	32	24	28
1/2	41	53	47
3/4	8	14	11
9/10	3	3	3
9/10+	0	0	0
1/2 or less	89	83	86
3/4 or more	11	17	14

It will be seen from Table XII that in about 85 per cent. of the cases was the amount of cumulus and cumulo-nimbus cloud 5 tenths or less. Assuming that, in most of the cases, there are also some amounts of energy-consuming or passive clouds present, it will be seen that the amount of energy-producing cumulus rarely exceeds 5 tenths, which is in good agreement with what the slice method indicates. In this connection it should be noted that the amounts of convective cloud reported by the ground observer will exceed those reported by the air observer by several tenths on account of the perspective effect.

19. Cloud Base

As was shown in paras. 2 and 10, the condensation level is obtained from the surface observation by assuming that the air ascends dry-adiabatically without appreciable mixing with the environment. The condensation levels thus evaluated are independent of the theories underlying the parcel method and the slice method.

The condensation levels have been evaluated from the TePhi-grams (the selected series) and compared with the actually observed cloud bases, and the results are shown in Tables XIII–XVI.

TABLE XIII

Comparison between Observed Cloud Base and Computed Condensation Level. Selected Series. All Cases included.

Computed condensation level (mb.)	Observed Cloud Base (mb.)													Total
	1000 990	980 970	960 950	940 930	920 910	900 890	880 870	860 850	840 830	820 810	800 790	780 770	760 750	
1000–990		9	10	2	3	1	2		1				1	29
980–970	3	**14**	12	7	3	5								44
960–950	1	4	**15**	15	9	6	4	1	2	1		1	1	60
940–930		1	4	**13**	19	16	7			3			1	64
920–910		2	3	2	**4**	8	7	1	3	1			1	32
900–890			1	3	1	**2**	3	4	6	1	1			22
880–870							**1**	4	5	2	1	1		14
860–850							1	**2**	4		2	1		10
840–830							1			1		1		3
Total ..	4	30	45	42	39	38	26	12	21	9	4	4	4	278

Mean Observed Cloud Base .. .. .. = 908 mb.
Mean Computed Condensation Level .. = 938 mb.
Difference .. .. = –30 mb.
Correlation Coefficient .. .. = 0·62

Note.—Figures in bold type indicate the diagonal.

TABLE XIV

Comparison between Observed Cloud Base and Computed Condensation Level. Selected Series ; only Cases with Showers, Hail and Thunderstorms are included.

Computed Condensation level (mb.)	Observed Cloud Base (mb.)													Total
	1000 990	980 970	960 950	940 930	920 910	900 890	880 870	860 850	840 830	820 810	800 790	780 770	760 750	
1000–990		2	5		1				1				1	10
980–970	1	**6**	4	3	1	3								18
960–950		3	**11**	7	6	1	2	1	2	1		1		35
940–930		1	2	**4**	7	2				2			1	19
920–910		2	2		**2**	2	4	1					1	14
900–890				1	1		1	1						4
880–870														0
860–850								1		1				2
Total ..	1	14	24	15	18	8	7	4	3	4	0	1	3	102

Mean Observed Cloud Base .. .. .. = 917 mb.
Mean Computed Condensation Level .. = 950 mb.
Difference .. .. = –33 mb.
Correlation Coefficient .. .. = 0·32

Note.—Figures in bold type indicate the diagonal.

TABLE XV

Comparison between Observed Cloud Base and Computed Condensation Level. Selected Series ; only Cases with Convective Strato-cumulus, without Precipitation and other low Cloud Forms are included.

Computed condensation level (mb.)	Observed Cloud Base (mb.)													Total
	1000 990	980 970	960 950	940 930	920 910	900 890	880 870	860 850	840 830	820 810	800 790	780 770	760 750	
1000–990		2	3				(2)							7
980–970	1	**3**	5	1	2									12
960–950		1	**3**	2	1	1	2						(1)	11
940–930				**1**	4	6	2							13
920–910			1		**1**	1			1					4
900–890						**1**		1						2
880–870								1			1			2
860–850														0
840–830							1					1		2
Total ..	1	6	12	4	8	9	7	2	1	0	1	1	1	53

Mean Observed Cloud Base .. .. .. = 918 mb.
Mean Computed Condensation Level .. = 945 mb.
Difference .. .. = –27 mb.
Correlation Coefficient .. .. = 0·72

Note.—Figures in bold type indicate the diagonal.

TABLE XVI

Comparison between Observed Cloud Base and Computed Condensation Level. Selected Series ; only Cases with Clouds of Cumulus Type without Precipitation and other Low Cloud Forms are included.

Computed condensation level (mb.)	Observed Cloud Base													Total
	1000 990	980 970	960 950	940 930	920 910	900 890	880 870	860 850	840 830	820 810	800 790	780 770	760 750	
1000–990		4	4		2	1								11
980–970		**3**	1	4									(1)	9
960–950	1	1	**1**	5	2	2	1							13
940–930			2	**4**	4	8	3			1				22
920–910			1	1	**2**	4	2			1				11
900–890				2	1	**2**	3	3	1	1				13
880–870						1	**3**	5	2		1			12
860–850						1	2	**3**	1	1	1			9
Total . .	1	8	9	16	11	19	14	11	4	4	2	0	1	100

Mean Observed Cloud Base . . . . . . = 902 mb.
Mean Computed Condensation Level . . = 924 mb.
Difference . . . . = –22 mb.
Correlation Coefficient . . . . = 0·74

Note.—Figures in bold type indicate the diagonal.

It will be seen from Table XIII that there is a fair correlation between the computed condensation level and the observed cloud base ; but the scatter is appreciable, and there is a systematic difference of — 30 mb., showing that the cloud base is normally appreciably higher (approximately 900 ft.) than the evaluated condensation level.

An attempt has been made at grouping the observations according to type of cloud, weather, etc., and the results are shown in Tables XIV–XVI.

It will be seen from Table XIV that, when precipitation is falling from convective clouds, the correspondence is very slight, and there is a clear indication of the cloud base being above (approximately 1,000 ft.) the computed condensation level. The explanation for this is obvious. The precipitation causes the cloud base to become variable and largely unrelated to the surface conditions. Below the cloud base the air is moistened by the rain, and at the ground the air is moistened also by evaporation from terrestrial sources of water, with the result that the condensation level evaluated from surface observations will give too low values of the heights and too high values of pressures at the condensation level.

It will be seen from Table XV that, except for the three wild cases indicated by parentheses,* the correspondence is very good ; the correlation is 0·72, but there is a systematic difference of — 27 mb., showing that the base of convective strato-cumulus is appreciably higher (i.e. about 800 ft.) than the computed condensation level.

Table XVI shows that the correspondence between computed and observed base of cumulus clouds is very high ; the correlation is 0·74 and the mean difference is — 22 mb., or about 650 ft. It will be seen that the difference does not vary much with the height of the cloud base.

If there were any appreciable mixing between the ascending and descending masses below the cloud base one would expect that the difference between the observed and computed condensation level would increase with increasing distance from the ground to the cloud base, for the ascending air would then gradually mix with the descending air and become increasingly dryer as it ascends. Table XVI, however, shows no appreciable tendency in this direction.

The earth's surface being a source of moisture, it is to be expected that the air at the ground will normally contain more humidity than does the bulk of the air below the cloud base. If this is so, the computed condensation level would be too low, for a non-representative saturated-adiabatic would be used in evaluating the condensation level.

During the period to which the observations refer, no detailed observations of temperature and humidity were made within the friction layer, but such observations were initiated later. From 100 ascents made in convective situations at Aldergrove and Bircham Newton in July, 1944, the following mean values of the potential wet-bulb temperature were found.

Level	Potential wet-bulb temperature
(ft.)	(° F.)
0	56·6
500	55·3
1,000	54·8
1,500	54·6

It will thus be seen that the normal state in convective situations is characterised by an excess humidity in a relatively shallow layer near the ground and a gradual approach to a representative potential wet-bulb temperature at about 1,000 ft. above the ground. Using now the representative value 54·6 instead of the surface value 56·6, the condensation level is increased by a height corresponding to a little more than 20 mb., which is sufficient to account for the differences indicated in Tables XV and XVI.

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* These cases, as well as the one in Table XVI have been disregarded in the computation of the correlation coefficients and the means.

Let CB and CL denote the cloud base and the computed condensation level, both expressed in terms of pressures. Then, approximately

$$CB = CL - 25$$

If this relation is used in connection with the observations in Tables XV and XVI, one would obtain the cloud base with an accuracy of ± 25 mb. in about 75 per cent. of the cases ; errors of ± 50 mb. would occur only in about 12 per cent. of the cases.

20. Cloud Top

It was shown in paras. 3 and 4 that the parcel method leads to the result that the top of the convective cloud should reach to the top of the upper negative area, i.e. to the level G in Fig. 2. Poulter (1938), on the other hand, finds that the convective clouds tend to build up to the top of the positive area, i.e. to the level E in Fig. 2. Other authors contend that nature prefers a compromise between these two levels, which means that the top of the convective cloud would normally be found somewhere above the top of the positive area and below the top of the upper negative area.

It was shown in para. 8 that the slice method leads to the definite conclusion that the top of the convective cloud would only reach slightly beyond the level D in Fig. 2, i.e. the level where the lapse rate becomes equal to the saturated-adiabatic rate.

Now, if there is active convection below an inversion, the three levels D, E and G in Fig. 2 will approach one another, and the distance from D to E and from E to G will be relatively small. This is almost invariably so in the cases of convective strato-cumulus. In these cases, it is difficult to decide which of the theories renders the most satisfactory results. On the other hand, when convection results in cumulus, or cumulo-nimbus, the distances from D to E and from E to G will be very large, and a good check on the validity of the theories can be obtained.

Before we proceed to compare the observed heights of the cloud tops with the computed heights, it is necessary to remark that the ascents under investigation terminate at the 400 mb. level, with the result that the levels referred to may escape observation. This is often true of the level G (Fig. 2), which is often found above the 400 mb. surface. When this is the case, the levels are tabulated as being at <400 mb., indicating that the level in question is above the 400 mb. surface. The results of these comparisons are given in the tables below.

TABLE XVII

Percentage Frequencies of Observed Heights of Tops of Cumulus and Cumulo-nimbus and of Similar Heights computed by the Aid of the Slice Method and the Parcel Method. Selected Series. Only Cases of Cu and Cb are included

Pressure (mb.)	Observed cloud top	Computed height of cloud top		
		Slice method, i.e. level D	Parcel method, top of upper negative area	Parcel method, top of positive area
<400	—	—	47	24
400–440	—	—	5	6
450–490	2	2	5	4
500–540	2	1	5	2
550–590	1	4	5	5
600–640	3	2	5	5
650–690	8	2	6	11
700–740	25	6	11	11
750–790	19	28	5	11
800–840	19	23	3	11
850–890	15	20	1	10
900–940	6	12	2	—
Mean	765	790	<525	<625
Difference	0	−25	>240	>140

It will be seen from Table XVII that the results obtained by the slice method are entirely satisfactory. On the average, the pressure at the level D is 25 mb. higher than that at the top of the cloud, which agrees with the findings of the slice method, for the cloud was to penetrate into the stable layer slightly above the level D. In comparison it will be seen that the parcel method renders results that are greatly at variance with the observations. The top of the upper negative area is normally much higher than the top of the convective cloud, and so is also the top of the positive area.

To obtain a more reliable assessment of the accuracies rendered by the methods referred to above, the frequencies of differences in pressure, corresponding to Table XVII, are given in Table XVIII.

TABLE XVIII

Percentage Frequency of Pressure Difference between Computed and Observed Heights of Cloud Tops. Selected Series. Only Cases of Cu and Cb are included

Pressure difference (mb.)	Level D minus cloud top	Level G minus cloud top	Level E minus cloud top
< -200	—	61	29
-200 to -110	—	17	16
-100 to $-$ 60	5	10	14
$-$ 50 to $-$ 10	15	7	17
0	20	1	7
10 to 50	47	1	13
60 to 100	6	3	3
110 to 200	7	—	1
> 200	—	—	—
Mean	25	<-240	<-140

It will be seen from Table XVIII that the pressure differences resulting from the parcel method are large on the average, and their distribution is erratic. The slice method, on the other hand, gives a pronounced concentration of the frequencies at about 25 mb. Let p_D be the pressure at the level D, and p_c the pressure at the top of the cloud. Then, on the average, the slice method gives

$$p_C = p_D - 25 \text{ (mb.)}$$

If this formula were used to evaluate the heights of the tops of cumulus and cumulo-nimbus, an accuracy of ±25 mb. would be obtained in about 55 per cent. of the cases, and an accuracy of ±50 mb. would be obtained in about 80 per cent. of the cases.

The above results refer to cumulus and cumulo-nimbus clouds. In the case of convective strato-cumulus, there is normally an inversion present in the upper part of the cloud layer. The level D (see Fig. 2) will then normally coincide with the base of the inversion, and the top of the positive area will normally be within the inversion layer. Since, in these cases, the positive area below the inversion is relatively small, the upper negative area will also be small, with the result that the parcel method and the slice method give results that do not differ greatly. Normally, the strato-cumulus layer reaches 20–30 mb. into the inversion layer.

21. **Release of Convection : The Maximum Negative Area**

It was shown in para. 5 that an initial vertical velocity of 7·5 m/sec. is required for a unit mass to overcome a negative area of 1 cm² on the TePhi-gram, and even if the negative area of the surface air were as small as 0·18 cm² an initial velocity of 1 m/sec. would be required to overcome the buoyancy forces. From this it will be seen that the negative area of the surface air must be reduced to an immeasurable size before energy-producing convective currents can be released.

On the other hand, the horizontal temperature distribution is not uniform, and when the representative observations show a small negative area, the negative area is likely to vanish locally, and a positive area may be present in places. Thus, for any given negative area on an ascent curve there is a certain probability for a slight positive area to be present locally. Therefore, as pointed out in paras. 5 and 17, convection is likely to be released as soon as the negative area of the ascent curve reaches a certain value. We shall now endeavour to determine the maximum size of the negative area that will allow convection to be released. To this end, the selected series will be used. The problem is a statistical one, and our aim is to determine the maximum size of the negative area, such that when this value is used, satisfactory results will be obtained in at least 90 per cent. of the cases. In view of the discussion in para. 12 it seems reasonable to allow a margin of 10 per cent. to allow for " wild cases " arising from observational errors.

The cumulus, cumulo-nimbus, and strato-cumulus clouds in the selected series have been correlated with the positive and negative areas, and the results are shown in Table XIX.

It will be seen that the clouds fall into two categories, viz. :—

A. A group (representing about 12 per cent. of the total number) in the upper right-hand corner of the table, which is characterised by a very large negative area and, except in two cases, no positive area at all.

B. A group (representing about 88 per cent. of the total number) in the left-hand part of the table which, on the whole, is characterised by the absence of, or a very small, negative area.

Of the 35 cases in category A, 15 cases occurred with precipitation and a stable layer, or inversion, near the ground ; in 7 cases strato-cumulus was reported, and in the remainder of the cases, feeble fronts appeared to be present aloft. In all but 3 of the 35 cases, pronounced instability was found in the upper atmosphere. It is evident that the clouds referred to in category A resulted from convection released in the free atmosphere, and not from the ground. These cases will therefore be excluded from the following discussion.

Having removed the 35 cases in category A, there remain 250 cases to be examined.

TABLE XIX

Percentage Frequency of Cloud in Combinations of Positive and Negative Areas.

Selected Series.

Positive Area (cm²)	Negative Area (cm²)													Total
	0	0·1–0·5	0·6–1·0	1·1–2·0	2·1–3·0	3·1–4·0	4·1–5	6–10	11–20	21–30	31–40	41–50	>50	
0				0·3	0·3			0·3	3·2	1·8	2·1	1·1	3·2	12·3
0·1–0·5	3·9	3·9	2·1	1·1	0·3	0·3					0·3		0·3	12·3
0·6–1·0	2·5	3·2	1·4	0·7		0·3								8·1
1·1–2·0	8·4	3·2	1·8	0·3										13·7
2·1–3·0	4·9	1·1		0·3										6·3
3·1–4·0	6·0	0·3	0·7			0·3								7·4
4·1–5	3·2	1·4												4·6
6–10	10·5	2·1		0·3	0·7									13·7
11–20	12·3	2·1	1·4	0·3										16·1
21–30	3·9			0·3										4·2
31–40	1·1													1·1
41–50														—
>50	0·3													0·3
Total ..	56·8	17·2	7·4	3·9	1·4	1·1	0	0·3	3·2	1·8	2·5	1·1	3·5	100
No. of cases	Category B : 250							Category A : 35						285

When precipitation falls, the surface temperature and wet-bulb temperature will often become non-representative as a result of evaporation from terrestrial sources of water. We shall, therefore, discuss the cases with and without precipitation separately. The cases with precipitation are shown in Table XX.

TABLE XX

Percentage Frequency of Cloud in Combinations of Positive and Negative Areas. Selected Series. Only Cases with Precipitation are included.

Positive Area (cm²)	Negative Area (cm.²)							Total
	0	0·1–0·5	0·6–1·0	1·1–2·0	2·1–3·0	3·1–4·0	4·1–5·0	
0				0·4	0·4			0·8
0·1–0·5	2·0		1·2	0·4		0·4		4·0
0·6–1·0	1·6	1·2	0·4	0·4				3·6
1·1–2·0	2·4	1·2	0·8	0·4				4·8
2·1–3·0	0·8	0·8						1·6
3·1–4·0	1·6	0·4	0·4					2·4
4·1–5	0·8	0·8						1·6
6–10	4·4	1·6		0·4	0·4			6·8
11–20	6·8	1·2	1·2	0·4				9·6
21–30	2·4			0·4				2·8
31–40	1·2							1·2
41–50								
>50								
Total ..	24·0	7·2	4·0	2·8	0·8	0·4		39·2

Note.—The percentage refers to the total number 250 referred to in Table XIX.

TABLE XXI

Percentage Frequency of Cloud in Combinations of Positive and Negative Areas. Selected Series. Only Cases with Strato-cumulus without Precipitation and other Low Clouds are included.

Positive Area (cm²)	Negative Area (cm²)							Total
	0	0·1–0·5	0·6–1·0	1·1–2·0	2·1–3·0	3·1–4·0	4·1–5·0	
0								
0·1–0·5	0·8	2·8	1·4	0·4				5·4
0·6–1·0	1·6	2·0	1·6					5·2
1·1–2·0	4·4	1·2	0·8					6·4
2·1–3·0	0·8							0·8
3·1–4·0	2·4							2·4
4·1–5	1·2	0·4						1·6
6–10	1·2							1·2
11–20	1·2		0·4					1·6
21–30								
31–40								
41–50								
>50								
Total ..	13·6	6·4	4·2	0·4				24·6

Note.—The percentage refers to the total number 250 referred to in Table XIX.

TABLE XXII

Percentage Frequency of Cloud in Combinations of Positive and Negative Areas. Selected Series. Only Cases with Cumulus without Precipitation and other Low Clouds are included.

Positive Area (cm²)	Negative Area (cm²)							Total
	0	0·1–0·5	0·6–1·0	1·1–2·0	2·1–3·0	3·1–4·0	4·1–5·0	
0								
0·1–0·5	1·2	0·4	0·4					2·0
0·6–1·0	0·4			0·4				0·8
1·1–2·0	3·6	0·4						4·0
2·1–3·0	2·8	0·4		0·4				3·6
3·1–4·0	2·0		0·4			0·4		2·8
4·1–5		0·4						0·4
6–10	5·6	0·8						6·4
11–20	4·4	0·4	0·4					5·2
21–30	1·6							1·6
31–40	0·8							0·8
41–50								—
>50	0·4							0·4
Total ..	22·8	2·8	1·2	0·8		0·4		28·0

Note.—The percentage refers to the total number 250 referred to in Table XIX.

In order to obtain a reliable assessment of the maximum negative area compatible with convection, we shall eliminate the cases of mixed skies, and discuss separately the cases of convective strato-cumulus and cumulus. Table XXI refers to the cases of convective strato-cumulus without other low cloud forms, and Table XXII refers to the cases when cumulus and no other low cloud forms were observed.

(1) *Precipitation.*—It will be seen from Table XX that, in the majority of the cases, there is no negative area, but there is a fair number of cases with relatively large negative areas ; in about 20 per cent. of the cases is the negative area larger than 0·5 cm^2.

(2) *Convective strato-cumulus.*—Here, too (see Table XXI), the negative area is most frequently absent ; in about 19 per cent. of the cases is the negative area larger than 0·5 cm^2. and in less than 2 per cent. of the cases does the negative area exceed 1 cm^2.

(3) *Cumulus.*—Table XXII shows a much greater preference for the small negative areas than do the other tables. It will be seen that in about 81 per cent. of the cases is there no negative area, and in 91 per cent. is the negative area 0·5 cm^2 or less.

Allowing a margin of about 10 per cent. for observational errors, it seems safe to conclude that *cumulus clouds resulting from convection from the ground are not likely to form until the negative area has been reduced to about* 0·5 *cm*^{2}.

It will be seen from Table XXI that, within the same margin of error, convective strato-cumulus may be present even when the negative area slightly exceeds 0·5 cm^2.

Adopting 0·5 cm^2. as a convenient upper limit of the negative area that is compatible with convection from the ground, and making no allowance for observational errors, the above tables indicate the following success :—

	Per cent.
All cases	84
Convection with precipitation	80
Sc without precipitation and other low clouds	81
Cu, or Cb without precipitation and other low clouds	91

22. **Release of Convection : Diurnal Heating**

We shall now return to the discussion in para. 10 and try to determine the success that can be obtained in the forecasting of convective clouds due to diurnal heating by evaluating the convection temperature T$_c$, the maximum temperature T$_M$, and the corresponding condensation levels B and B′, as shown in Fig. 6.

To test the methods suggested in para. 10, the selected series of observations has been used. In order to eliminate all cases of advective heating or cooling, only ascents satisfying the following conditions were used :—

1. No convective cloud present at the time of the morning ascent, but cumulus or cumulo-nimbus present at the time of the midday ascent (i.e. 1200 G.M.T.).
2. Wind at station level force 3, or less.
3. No appreciable change in the temperature at and above the 800 mb. surface from morning to midday.

Only 32 ascents were found that satisfied all these requirements. Although the number is small, the results ought to be highly representative of diurnal convection.

For the evaluation of the convective temperature and the condensation level, the surface humidity is of great importance. In evaluating the representative humidity, the method suggested in para. 10 was used, i.e. the mean value through the layer up to the 950 mb. surface was evaluated from the TePhi-grams. By the aid of this value the condensation level (B) and the convection temperature (T$_c$) (see Fig. 6) were evaluated.

Next, the appropriate areas (A) given in Table D (para. 10) were applied to the morning ascents, and the maximum temperature (T$_M$) and the condensation level (B′) (see Fig. 6) were evaluated. Finally, the surface temperature (T$_o$), the observed cloud base (B$_o$) at the time of the midday ascent, and the area (A$_o$) gained from the morning to the midday ascent were evaluated and compared with the theoretical values.
The ratio of the observed area to that given in Table D is given in Table XXIII.

TABLE XXIII

Percentage Frequency of the Ratio A_o/A

Ratio	0·1	0·2	0·3	0·4	0·5	0·6	0·7	0·8
Frequency	6	25	16	19	13	12	3	6

Mean ratio 0·4

It will be seen that the observed areas are much smaller than the values indicated in Table D. It should be noted that the areas in Table D refer to the time of maximum temperature, and should, therefore, on the average, be larger than the areas observed at the time of the midday ascent, i.e. 1200 G.M.T. Nevertheless, the ratios given in Table XXIII seem to suggest that the areas given in Table D are definitely too large, since more than 40 per cent. of the available energy must have been supplied by 1200 local time.

We shall next consider the evaluated temperatures T_M and T_c. Table XXIV shows the frequencies of the difference between the temperatures observed at 1200 G.M.T. and T_M and T_c, respectively.

TABLE XXIV

Percentage Frequency of Differences $T_o - T_M$ *and* $T_o - T_c$

Difference (° F.)	$T_o - T_M$	$T_o - T_c$
<−9		6
−9 and −8		
−7 and −6	6	3
−5 and −4	25	6
−3 and −2	22	19
−1 to +1	25	16
2 and 3	13	13
4 and 5	9	22
6 and 7		9
8 and 9		
>9		6
Mean	−4	+2

It will be seen that the maximum temperatures, evaluated as described in para. 10, are about 4° F. higher than the temperatures observed at 1200. This difference is altogether reasonable, and since the range of the differences $T_o - T_M$ is fairly small, it appears that Gold's method of forecasting the maximum temperature gives results that are largely satisfactory.

The fact that Gold's method renders satisfactory values for T_M while the areas used are too large, can be explained by reference to Fig. 6. Thus, if the lapse rate above the ground is super-adiabatic, instead of adiabatic, a smaller area would give the same value for T_M. Furthermore, if the line MM in Fig. 6 is moved a distance corresponding to 1° F., the area will vary in proportion to its size. Thus, T_M becomes rather insensitive to an error in the area when the area is large, but it is highly sensitive to such errors when the area is small.

It will be seen from Table XXIV that, on the average, the temperature observed at 1200 G.M.T. is about 2° F. higher than T_c. This is what one would expect, since convection was actually released by 1200 G.M.T. However, in about 35 per cent. of the cases, convection had been released although the surface temperature had not reached the critical value. The reason for this may, in part, be due to the circumstance that convection is released over over-heated localities before the representative temperature reaches the critical value. This question was discussed in some detail in para. 21.

It will be seen from Table XXIV that the range of the variation in $T_o - T_c$ is appreciable. The extreme cases were found to occur when the ascent curve was irregular near the earth's surface. Furthermore, when the lapse rate in the morning is such that the temperature curve is parallel to the lines of humidity mixing ratio, the intersection is poor, and great errors in the evaluated T_c will result.

It will be seen from Fig. 6 that the point B represents the level at which convective clouds would begin to form, assuming that the humidity mixing ratio remains sensibly constant. On the same assumption, and using Gold's areas, the condensation level at the time of maximum temperature would be found at B′. The levels B and B′ have been evaluated and compared with the height of the base of the convective cloud, and the results are shown in Tables XXV and XXVI, where B_o refers to the observed cloud base.

TABLE XXV

Percentage Frequencies of $B_o - B'$

Difference (mb.)	Cloud base (mb.)			
	Below 950	950–910	900–860	Above 860
<−60				
−60 and −50				3
−40 and −30				3
−20 and −10			9	
0			6	
10 and 20		3	6	
30 and 40		6	9	
50 and 60	3	9	6	3
>60	22	9		

Mean without regard to sign .. .. 48 mb.
True mean .. .. 39 mb.

It will be seen from Table XXV that the condensation levels obtained by the aid of the areas in Table D are, on the average, much too high, i.e. they occur about 40 mb. above the observed cloud base. This suggests that the areas in Table D are too large.

TABLE XXVI

Percentage Frequencies of $B_o - B$.

Difference (mb.)	Cloud base (mb.)			
	Below 950	950–910	900–860	Above 860
<-60			6	6
-60 and -50			3	
-40 and -30		3	13	
-20 and -10	3	6		
0		3	3	
10 and 20	9	3	3	
30 and 40	3	6	3	3
50 and 60		6		
>60	9		6	

Mean without regard to sign .. .. 44 mb.
True mean .. .. 6 mb.

It will also be seen that the error is large (i.e. > 60 mb.) when the cloud base is low, and it decreases as the height of the cloud base increases. The reason for this can be explained by reference to Fig. 6. Suppose that the ascent in the morning showed a ground inversion. The point A would then be to the left of the position indicated, and there would be a maximum of temperature at some distance above the ground. To compensate for the area due to the ground inversion, the line MM would only have to be moved slightly downwards, with hardly any effect on the level B′. Thus, the level B′ is rather insensitive to the structure of the air above the ground.

Suppose next that the humidity mixing ratio were considerably higher than indicated in Fig. 6. The line CB would then be found further to the right. But when this occurs, the distance from B to B′ would increase rapidly, depending upon the lapse rate in the air column. Thus, when the condensation level is low, the level B′ will normally be much too high. If, on the other hand, the humidity were less than that indicated in Fig. 6, the line CB would be found further to the left, and the level B′ would be representative.

From the above discussion, it follows that the area method is not suitable for evaluating the condensation level of convective clouds.

It will be seen from Table XXVI that, on the average, the level B agrees well with the observed cloud base, but the scatter is very large, and the error varies with the height of the cloud base. Thus, neither of the two methods of predicting the cloud base at noon by the aid of the morning ascents, give satisfactory results. The errors in the level B are mainly due to the complexities of the ascent curves, and to the poor intersections that are obtained when the ascent curve AB becomes parallel to the humidity curve CB. Appreciable errors are also found when the humidity distribution along the vertical is irregular, and also when the ground is so wet that the dew point varies appreciably as the heating progresses.

To improve on the methods of forecasting the condensation level, the convective temperature and the maximum temperature, more accurate assessments of the available energy, and of the transfer along the vertical of heat and moisture, would be needed.

23. **Equivalent Cloud Amount**

It was shown in para. 21 that whether or not convection will occur depends upon the size of the negative area. The size of the positive area appears to be related to the amount of convective cloud, but this relationship is not apparent if the positive areas are compared with the cloud cover, i.e. the number of tenths of the sky covered by cloud, for when the positive area is small, there is usually an inversion above the unstable layer, and convection will then tend to produce a thin layer of cloud that covers a large portion of the sky. However, if the cloud cover and the depth of the cloud are converted into equivalent cloud amounts, as explained in para. 17, the relationship shown in Table XXVII is found.

TABLE XXVII

Equivalent Cloud Amounts. Non-selected Series.
Only Cases with Cumulus or Cumulo-nimbus are included.

Positive Area (cm^2)	Mean Equivalent Cloud Amount	Number of Cases
0·1– 1·0	3	35
1·1– 3·0	4	26
3·1–10·0	9	32
11·0–20·0	19	16
>20	13	10

It will be seen from Table XXVII that, on the average, there is an almost linear relation between the size of the positive area and the equivalent cloud amount, but the scatter of the individual values around the mean has been found to be appreciable.

24. Release of Precipitation

As mentioned in para. 13 there is reason to believe that the presence of ice-crystals and super-cooled water in the clouds is a necessary condition for the release of precipitation of the intensity that is normally observed in connection with convective clouds*. It was also pointed out that the depth of the cloud is an important factor.

In the selected series, 99 cases of precipitation occurred where the temperature at the top of the cloud and the depth of the cloud layer could be evaluated. Of these, 6 observations showed light showers with temperature above freezing at the top of the cloud. In 5 cases light drizzle was reported from strato-cumulus with temperature above freezing at the top of the cloud. In 2 cases it was doubtful whether the light rain reported fell from convective clouds or from an alto-stratus deck present at higher levels.

The remaining 86 cases, in which the temperature at the top of the cloud was below freezing, have been grouped according to temperature intervals and depth of the cloud layer, and the results are shown in Table XXVIII.

TABLE XXVIII

Percentage Probability of Precipitation in Combinations of Temperature at the Top of the Cloud and Depth of Cloud Layer. Selected Series.

Depth of Cloud Layer (mb.)	Temp. (°F.) at Top of Cloud				Number of Cases with Precipitation	Number of Cases without Precipitation	Probability Per cent.
	32–28	27–23	22–18	<18			
<50	14	40	0	40	4	20	17
60–100	1	17	50	29	6	22	21
110–150	0	29	40	29	6	18	25
160–200	13	22	60	36	10	24	29
210–300	33	30	75	59	17	17	50
>300	33	—	—	82	43	11	80
Number of Cases with Precipitation.	6	10	10	60	86		
Number of Cases without Precipitation.	39	27	11	35		112	
Probability, per cent.	13	27	48	63			43

The number of cases in each square of the main body of the table is too small for reliable probabilities to be obtained, but the probabilities in the right-hand column and in the bottom row contain a sufficient number of cases to give significant values. It will be seen that the probability of release of precipitation increases uniformly with decreasing temperature below freezing. It will also be seen that the probability increases rapidly with increasing cloud depth, and when the cloud depth exceeds 300 mb., the probability of precipitation is 80 per cent. It appears, therefore, that a large cloud depth is at least as important as is a sub-freezing temperature in the cloud. It should be emphasized that the probabilities indicated above refer to the occurrence of precipitation at the time of observation. If the temperatures and the cloud depths had been correlated with the occurrence of precipitation within a time interval, say three hours, substantially higher probabilities might have been found.

From the above discussion it follows that observations seem to substantiate Bergeron's theory for colloidal instability. The few cases, mentioned at the beginning of this paragraph, in which precipitation fell from clouds without the temperature in the cloud being below freezing, do not necessarily contradict the theory, for drops of drizzle dimensions are normally released without the assistance of ice crystals.

25. The Lapse Rate within Convective Clouds

The selected series has been used to evaluate the lapse rate that normally occurs within convective clouds. Observations are made at intervals of 50 mb. The evaluated lapse rate will, therefore, represent mean values through relatively deep layers. Particularly when the convective currents are strong, the observing aircraft prefers to fly between the clouds, with the result that the cases of ascents through convective clouds are relatively rare, and will not contain cases of intense convection. On account of this, and because of the smoothing over 50 mb. layers, lapse rates considerably larger than those indicated below are likely to occur.

It is convenient to compare the lapse rate within the cloud with the saturated-adiabatic rate at the appropriate pressure and temperature. In the notations used in the foregoing paragraphs, the *excess lapse rate* is defined as

$$\gamma - \gamma_s$$

Since the saturated-adiabatic lapse rate varies with pressure and temperature while the dry-adiabatic lapse rate is constant, it is convenient to express the excess lapse rate as a percentage of the dry-adiabatic rate. Thus

$$\gamma_E = 100 \cdot \frac{\gamma - \gamma_s}{\gamma_d}$$

When the excess lapse rate is positive the cloud is unstable, and when it is negative the cloud is stable.

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* It is generally recognised that this condition need not be satisfied for the release of droplets of the drizzle size.

The excess lapse rates (γ_E) derived from the selected series, are given in Table XXIX.

TABLE XXIX

Percentage Frequency of Excess Lapse Rates expressed as a Percentage of the Dry-adiabatic Rate. Selected Series. Total Number of Cases 63.

γ_E (per cent.)	Pressure (mb.)			Total
	>800	800–600	600–400	
<−20	1·5	3·0	1·5	6·1
−20 to −11	3·0	1·5	1·5	6·1
−10 to − 1	1·5	1·5	3·0	6·1
0	7·6	10·6	3·0	21·2
1 to 10	9·1	3·0	3·0	15·2
11 to 20	6·1	9·1	15·1	30·2
21 to 30	9·1	1·5	—	10·6
>30	3·0	1·5	—	4·5
Total ..	40·9	31·8	27·2	100·0
Stable ..	6·1	6·1	6·1	18·3
Neutral ..	7·6	10·6	3·0	21·2
Unstable ..	27·3	15·1	18·1	60·5

It will be seen from Table XXIX that instability within the cloud mass is a very frequent occurrence. It is of interest to note that in 5 per cent. of the cases was the lapse rate within the cloud equal to the dry-adiabatic rate, and in 10 per cent. of the cases did it exceed that rate. The largest excess lapse rate was 43 per cent. of the dry-adiabatic rate.

26. **Vertical Velocity and Gusts**

We shall now endeavour to assess the order of magnitude of the vertical velocities and accelerations in the convective currents by applying the theories of convection to the observed temperatures, lapse rates and humidities. The accuracy that can thus be obtained will depend greatly on the underlying assumptions. We shall apply the theory first to saturated ascent through a dry environment (i.e. cloud versus cloudless air), and then to saturated ascent through a saturated environment (i.e. vertical gusts within the cloud).

(1) *Saturated ascent through a dry environment.*—As shown in paras. 2 and 3, the parcel method leads to the result that the maximum vertical acceleration occurs at the level D in Fig. 2, and the maximum vertical velocity at the level E. It was shown in para. 5 that the vertical velocity at the level E would be proportional to the square root of the positive area below the level E. These areas have been evaluated from the ascent curves and converted into velocities, using Table C in para. 5. The results are shown in Table XXX.

TABLE XXX

Percentage Frequency of Computed Vertical Velocities at the Top of the Positive Area, or at the 400 mb. Level if the Positive Area extends beyond that Level. Selected Series.

W (m/sec.) (computed)	Clouds of cumulus type	Strato-cumulus
0·1–1·0	—	—
1·1–2·0	1	3
2·1–3·0	2	7
3·1–4·0	4	5
4·1–5·0	3	7
5·1–10	19	47
11–15	17	13
16–20	13	8
21–25	13	7
26–30	13	3
31–40	10	—
41–50	4	—
>50	1	—
Total ..	100	100
Mean Velocity ..	19 m/sec.	10 m/sec.

It will be seen that these velocities bear no resemblance to observed facts. Thus, according to Table XXX, the mean vertical velocity at the top of the positive area would be about 20 m/sec. in cumulus situations, and 10 m/sec. in strato-cumulus situations. These velocities would represent the mean values for the entire mass of saturated air. On the other hand, it was shown in para. 7 that the parcel method, when applied to saturated ascent through a dry environment, leads to a very great overestimate of the available energy, and, hence, the velocities given in Table XXX will be much too large.

It was shown in para. 8 that the ascending velocity of the saturated air as a whole, relative to the non-saturated environment, must normally be considerably less than 1 m/sec., and it was shown in para. 20 that the vertical velocity and acceleration of the ascending cloud mass are usually nullified more than 100 mb. below the level E in Fig. 2, where the parcel method leads to mean vertical velocities of gale force. In view of this, the values given in Table XXX must be discarded as having no foundation in re.*

(2) *Vertical gusts within the cloud.*—It was shown in paras. 6, 7 and 9 that the parcel method leads to an underestimate of the available energy when both the ascending and the descending masses follow the same adiabatic. The under-estimate of the available energy is proportional to the factor M′/M, and if the ascending and the descending masses are about equal, the parcel method will give only about 50 per cent. of the true amount of energy.

Below the cloud base the lapse rate is only slightly super-adiabatic through a relatively shallow layer, and the positive area will be small, with the result that the vertical gusts are not excessive.

Within the cloud, the conditions are different, for when the rising air reaches the condensation level, with a lapse rate which is almost dry-adiabatic, a very large positive area becomes immediately available and strong gusts will occur. If no mixing occurred between the ascending bubbles and their environment, excessive velocities would result. Mixing, however, will smooth out the temperature contrasts, and the velocities that result will depend largely upon the lapse rate within the cloud and the life of the bubbles.

To obtain an assessment of the vertical gusts within the cloud, we shall use the lapse rates given in Table XXIX. We shall assume that the ascending bubble within the cloud is sufficiently large to maintain its identity while it ascends through a layer 50 mb. deep.

As pointed out in para. 25, the lapse rates given in Table XXIX do not represent maximum values, for (1) they have been smoothed over a 50 mb. layer, and (2) the observing aircraft will avoid extreme conditions.

The lapse rates given in Table XXIX have been converted into the accelerations and velocities that would obtain when a bubble has moved through a layer 50 mb. deep. The results are shown in Tables XXXI and XXXII.

TABLE XXXI

Percentage Frequency of Vertical Accelerations after Passage through a 50 mb. Layer, derived from the Lapse Rates in Table XXIX. Only Cases of Instability are included.

Acceleration (m/sec.^{2})	Level (mb.)			Total
	Below 800	800–600	600–400	
0·01	10·0	2·5	—	12·5
0·02	10·0	7·5	5·0	22·5
0·03	5·0	5·0	2·5	12·5
0·04	15·0	5·0	22·5	42·5
0·05	—	2·5	—	2·5
0·06	5·0	2·5	—	7·5
Total ..	45·0	25·0	30·0	100·0

It will be seen that the accelerations are very small. Even if the bubble ascended through a 100 mb. layer without loss of excess temperature, the accelerations would only be twice as large as indicated in the table, and if the excess lapse rate, in extreme cases, should be twice as large as the steepest one contained in Table XXIX, the acceleration would be twice as large as the largest one in the above table. It appears, therefore, that accelerations in excess of 0·25 m/sec.^{2} are extremely unlikely even in the most active cumulo-nimbus cloud.

With regard to the vertical velocities of the gusts within the cumulo-nimbus cloud, similar considerations hold. According to Table XXXII, vertical gusts of 6 m/sec. should be a fairly frequent occurrence in convective clouds that reach above the 600 mb. surface. Allowing, as above, for a longer life of the individual bubbles, and lapse rates that occasionally are steeper than the steepest one contained in Table XXIX, gust velocities in the vicinity of 20 m/sec. might be reached in extreme cases.

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* An attempt has been made at applying the theory of the so-called " virtual mass " to the convective elements. In its present state the theory applies only to solid bodies moving through an incompressible fluid, the velocity around the body being determined by a potential. The theory shows that the kinetic energy of the body is less than that which would result if the environment were undisturbed.

According to this theory the velocities in Table XXX should be multiplied by a factor less than unity, the value of the factor depending upon the shape of the body. Such factors have been computed for ellipsoidal bodies and it was found that the factor varies from 0·50 to 0·96 as the ellipticity varies from 1/5 to 5/1, with a factor of 0·85 appropriate to a spherical body.

It will be seen that even if these factors are applied, the velocities in Table XXX would still be far too large, showing that assumptions underlying the theory of the virtual mass has but little bearing on convection in the atmosphere.

TABLE XXXII

Percentage Frequency of Vertical Velocities after Passage through a 50 mb. Layer, derived from the Lapse Rates in Table XXIX. Only cases of Instability are included.

Velocity (m/sec.)	Level (mb.)			Total
	Below 800	800–600	600–400	
1				
2	10·0			10·0
3	10·0	7·5		17·5
4	15·0	7·5	7·5	30·0
5	10·0	7·5	5·0	22·5
6		2·5	17·5	20·0
Total ..	45·0	25·0	30·0	100·0

With regard to gust gradients within the convective cloud, no deductions can be made from the observational material examined.

27. **Relation to Synoptic Situation**

The occurrence of convection has been correlated to the synoptic situation, and some of the results obtained were described in para. 15, in order to characterize the selected series of observations. Some additional results are given in the tables below.

(1) *Vertical Shear.*—The vertical shear of the horizontal wind through the convective layer is usually small. On the average, the shear per 50 mb. layer from the base to the top of the cloud was found to be 2 km. per hour. The shear increases rapidly above the cloud top; the mean shear through the 50 mb. layer above the cloud top was found to be 8 km. per hour.

The shear is not uniformly distributed through the cloud layer. On the average, there is a slight shear at the cloud base; uniform velocity through the main cloud layer, and increasing shear towards the top of the cloud, with strong shear above the cloud top. Since strong wind shear is indicative of the presence of horizontal temperature gradients, the shear above the cloud suggests that the cloud tops reach up to the first stable layer, in agreement with what was found in paras. 8 and 20.

The frequencies of variation of wind direction through 50 mb. layers, from the base to the top of the convective cloud, are shown in Table XXXIII.

TABLE XXXIII

Percentage Frequency of Variation in Wind Direction: (1) Convection, (2) regardless of Convection, and (3) above Cloud Top.

Degrees	(1)	(2)	(3)
<−30	—	2	1
−30	1	1	1
−20	3	5	1
−10	13	15	20
0	65	46	54
10	14	22	20
20	3	5	3
30	1	2	—
>30	—	2	—
Max. ..	30	150	40

It will be seen from Table XXXIII that, in convective situations, the wind direction is very uniform along the vertical.

(2) *Vorticity.*—It was shown in para. 15 that there was a slight tendency for convection to occur in cyclonically curved sea level isobars. In an earlier paper (M.R.P. 207) it was found that there was a very high association between the occurrence of subsidence and anticyclonic vorticity in the free atmosphere. Since subsidence is a stabilising process, it would seem that convection might not readily occur together with anticyclonic vorticity. This point has been examined and the occurrence of convection has been correlated with the curvature and the shear of the flow in the 750 mb. surface. The results are given in Table XXXIV.

TABLE XXXIV

Percentage Association between the Occurrence of Convection and Vorticity at the 750 mb. Level. Selected Series. All Cases included.

Curvature	Shear			
	None	Cyclonic	Anticyclonic	Sum
None	19	12	4	35
Cyclonic ..	13	30	5	48
Anticyclonic ..	5	3	9	17
Sum ..	37	45	18	100

It will be seen from Table XXXIV that the association between convection and cyclonic vorticity is not very high. In 55 per cent. of the cases there is cyclonic vorticity, in 18 per cent. anticyclonic vorticity, in 8 per cent. the vorticity is indeterminate, and in 19 per cent. there is neither cyclonic nor anticyclonic vorticity.

We shall now divide the cases into three classes according to the height of the tops of the convective clouds, such that *shallow convection* will refer to the cases where the cloud tops were found at, or below, the 800 mb. surface, *deep convection* will refer to the cases where the cloud tops reached to, or above, the 600 mb. surface, and *very deep convection* where the cloud tops reached to, or above, the 400 mb. surface. Each of these groups has been correlated with the curvature and shear of the flow in the 750 mb. surface, and the results are shown in the tables below.

TABLE XXXV

Percentage Association between the Occurrence of Shallow Convection and Vorticity at the 750 *mb. Level. Selected Series. Number of Cases* 96.

Curvature	Shear			
	None	Cyclonic	Anticyclonic	Sum
None	20	7	5	32
Cyclonic ..	7	15	7	29
Anticyclonic ..	12	3	24	39
Sum ..	39	25	36	100

TABLE XXXVI

Percentage Association between the Occurrence of Deep Convection and Vorticity at the 750 *mb. Level. Selected Series. Number of Cases* 73.

Curvature	Shear			
	None	Cyclonic	Anticyclonic	Sum
None	13	11	1	25
Cyclonic ..	18	56	1	75
Anticyclonic ..	—	—	—	—
Sum ..	31	67	2	100

TABLE XXXVII

Percentage Association between the Occurrence of very Deep Convection and Vorticity at the 750 *mb. Level. Selected Series. Number of Cases* 31.

Curvature	Shear			
	None	Cyclonic	Anticyclonic	Sum
None	6	10	—	16
Cyclonic ..	13	71	—	84
Anticyclonic ..	—	—	—	—
Sum ..	19	81	—	100

It will be seen from Table XXXV that there is a slight preference for anticyclonic vorticity aloft when the convection is shallow, but the scatter is very great, and all possible combinations of shear and curvature may occur. On the other hand, Table XXXVI shows a very clear preference for cyclonic vorticity. Thus, in 85 per cent. of the cases is the vorticity cyclonic, in 1 per cent. anti-cyclonic, in 1 per cent. indeterminate, and in 13 per cent. there is no vorticity. The association is even higher in the case of very deep convection, for the vorticity is cyclonic in 94 per cent., and absent in the remaining 6 per cent. of the cases.

From the above it appears that deep and very deep convection should not be forecast unless the upper air charts indicate cyclonic vorticity.

28. Some Cases of Deep Convection

Some typical cases of deep convection are listed in Table XXXVIII, together with the observations of weather, cloud, etc. The mean values of temperature and wet-bulb temperature are plotted in Fig. 8.

It will be seen that the lapse rate above the ground is slightly super-adiabatic, i.e. there is no negative area. The mean cloud base is at about 925 mb. Below this level, the lapse rate of the wet-bulb temperature exceeds the saturated-adiabatic rate. The air below the cloud base is, therefore, convectively unstable. The same is true, to a lesser extent, from the cloud base to about 750 mb., above which level T_w follows a saturated-adiabatic. The lapse rate of temperature is, however, in excess of the saturated-adiabatic rate throughout, with the result that the cloud will be in an unstable state. Since there is no stable layer below the 500 mb. surface, the clouds will build beyond this level.

The convective instability in the lowest strata is an important factor, for when this air rises and becomes saturated, a super-adiabatic lapse rate will be maintained within the cloud, giving rise to strong convective gusts.

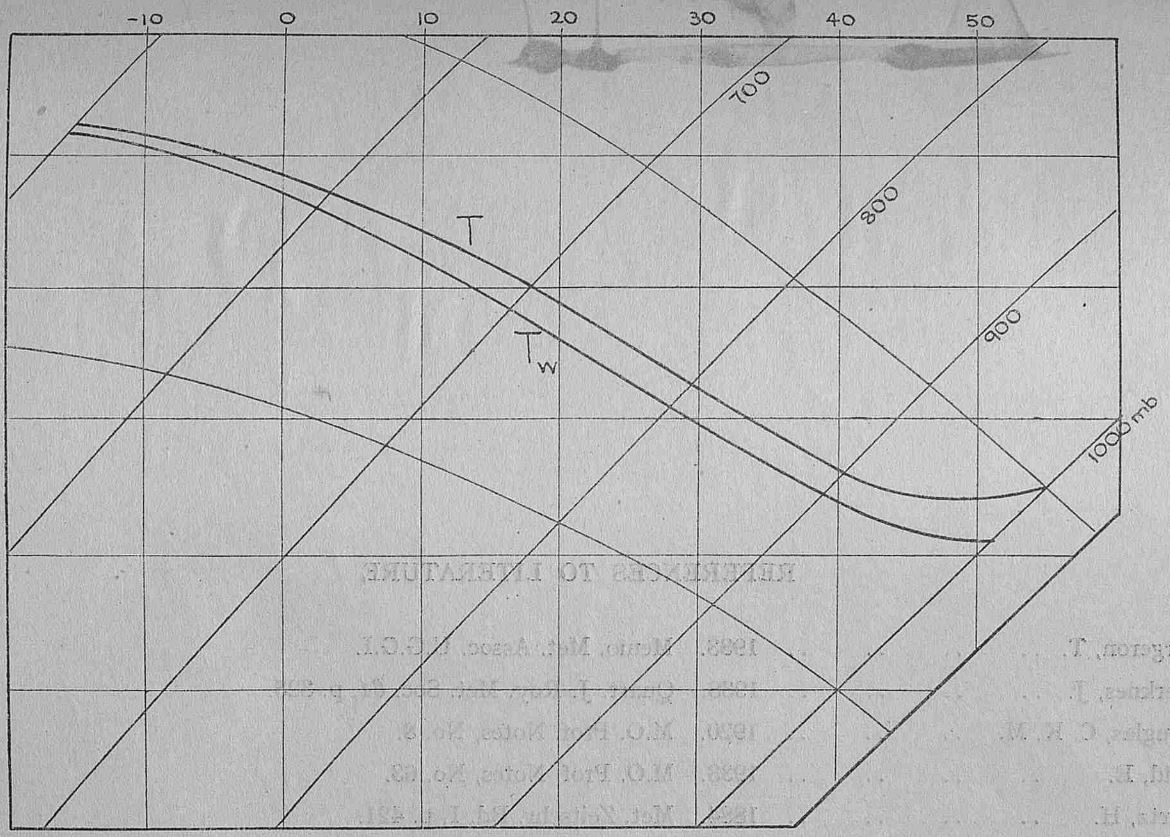

FIG. 8.—Mean temperature and wet-bulb temperature in 17 typical cases of deep convection. Two saturated-adiabatics on either side of the ascent curves, are indicated. The mean cloud base is at 925 mb. Note that the air is pronouncedly convectively unstable above the ground.

It will be seen from Fig. 8 that the wet-bulb depression is positive throughout. This is due to the circumstance that, in the majority of the cases, the observing aircraft ascended outside the cloudy columns. However, when the lapse rate outside the cloud is steep, the mean lapse rate within the cloud will also be steep, for, otherwise, excessive temperature contrasts would exist above the cloud tops. It follows then, that a steep lapse rate outside the cloud is indicative of an equally steep lapse rate within the cloud, and the chances for strong gusts to occur within the cloud can be estimated from the lapse rate observed outside the cloud.

Finally, it will be seen from Fig. 8, that the environment of the cloud is in a stable state, for its lapse rate is less than the dry-adiabatic. Within the cloud, on the other hand, the lapse rate is greater than the saturated-adiabatic, with the result that the cloud represents a *chute of instability* through the convective layer. A bubble that ascends from the ground and reaches the cloud base, will become energy-producing, whereas a bubble that ascends from the ground into the air between the cloudy column, will become energy-consuming. Hence, a convective circulation once started, will maintain itself and feed on the bubbles below the cloud while it tends to consume the energy of the bubbles that rise from the ground in the vicinity of the cloud. This state will continue as long as a super dry-adiabatic lapse rate is maintained above the ground. Hence to maintain convection, the upward transfer of heat through the cloud must be compensated for by a continued heating from the underlying surface. In these circumstances, a considerable portion of the energy areas given in Table D (para. 10) is lost to the upper strata, and this, at least in part, accounts for the absence of agreement between the theoretical and the observed energy areas referred to in para. 22.

TABLE XXXVIII

Typical Cases of Deep Convection. Aldergrove, 1943.

Date	Time	Cloud	Weather	Cloud						Negative Area (cm^2)	Positive Area (cm^2)
				Base (mb.)	Top (mb.)	Base (°F.)	Top* (°F.)	Amount (tenths)	Equivalent Amount		
May 9	12	Cb, Sc	Hail ..	950	<400	35	−47	5	>55	0	26
June 12	12	Cb, Cu	Rain showers	970	<400	50	−35	5	>57	0	>21
,, 12	18	Cb, Cu, Sc	Rain showers	970	<400	51	−31	3	>34	0	14
,, 14	18	Cb, Cu, Sc	Rain showers	940	470	46	−20	2–8	29	0	>10
,, 15	12	Cb, Cu	Past shower	960	<400	46	−41	3–6	>85	0·4	> 7
,, 16	12	Cb	Rain showers	940	490	45	−17	8	72	0	20
July 7	06	Cu, Sc	Rain showers	840	440	33	−28	8	64	0	> 3
,, 12	06	Cb, Sc	Nil ..	930	500	45	−15	5	43	0	>14
,, 12	12	Cb, Cu	Rain showers	870	<400	40	−34	5	>47	0	>16
,, 12	18	Cb, Cu	Rain showers	910	430	44	−29	7	67	0·1	>16
Aug. 20	12	Cb, Cu	Rain showers	940	410	58	−17	3–5	54	0	>16
,, 21	18	Cb, Cu, Sc	Rain showers	910	400	52	−25	6	54	0	>16
,, 25	18	Cb, Cu, Ac	Rain showers	850	<400	37	−36	5–8	>64	0·6	12
,, 26	12	Cb, Sc	Nil ..	920	500	46	−13	5	42	0·4	>14
,, 26	18	Cb, Cu, As	Past shower	950	420	48	−30	1–5	16	0	>15
Oct. 2	12	Cb, Cu	Rain showers	920	450	37	−30	2–5	23	0	14
Nov. 29	12	Cb, Sc, Ac	Past shower	960	440	36	−42	3–7	74	0·4	7

* Where the cloud reaches above the 400 mb. surface, the temperature refers to the 400 mb. surface.

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AIR MINISTRY
METEOROLOGICAL RESEARCH COMMITTEE

Problems of Forecasting in low latitudes

by F.K. Hare

Forecasting in low latitude presents special and acute difficulties. Most of the technique of forecasting in middle and high latitudes becomes useless near the equator. The essential steps in forecasting outside the equatorial belt can be summarised as follows:-

(i) identification of fronts and the prediction of their movement;

(ii) identification of airmasses and the prediction of their movement and modification; and

(iii) an estimate of the stability and any probable changes due to divergence.

Steps (i) and (ii) require for their completion an accurate analysis of the wind field at the surface and in the upper air. The difficulties of equatorial meteorology nearly all spring from the virtual impossibility of making such an analysis by any known method. Some of these difficulties are summarised below. An exact limit to the equatorial belt cannot be given, but it certainly includes the belt within the 10° parallels, and probably that within the 15° parallels.

1. The determination of the wind field.

This is the first step in analysis in all latitudes. In middle latitudes the wind is found by drawing up the pressure field, or by drawing contours of isobaric surfaces in the upper air. The derivation is based on the gradient wind equation, which is known to give a good approximation to the true wind at all levels. By this means the horizontal components of the wind are fairly accurately known, but the vertical component can only be guessed at. In low latitudes, however, the pressure gradient ceases to be a good measure of the wind, because of the decrease in the Coriolis parameter. The isobars can no longer be regarded as streamlines, and their spacing is no direct measure of the wind speed. Gradients are nearly always extremely gentle, isobars being few in number. This is the central problem of equatorial meteorology, i.e. that it is generally impossible to work out the wind distribution from any measurable physical property.

It is not at present known how far equatorwards the flow is along the isobars and at roughly the gradient speed. An immediate step in Tropical research must be to carry out an investigation analogous to Gold's "Barometric Gradient and Wind Force" (M.O.190) in latitudes within the 30° parallels.

It is also necessary to determine whether any approach to steady motion is possible in these latitudes. The semi-diurnal pressure wave keeps the pressure field in a permanently unsteady state, and it is difficult to see how steady circulation is possible.

It follows from the impossibility of determining the wind field that the divergence and vorticity are also hard to determine. The existence of wide rain areas in equatorial areas suggests that convergence and divergence exist on a large scale; it is therefore vital to be able to identify them. At present this is not possible; "divergence" is little more than a shibboleth used to explain the weather away. In middle latitudes the divergence is usually inferred rather than measured. The isallobaric gradient or the known divergent properties of certain isobaric patterns usually give a clue to it. Neither of these is of any value in low latitudes. A further clue now used in middle latitudes is bound up with the relationship between divergence and vorticity. The latter, however, cannot be calculated until the horizontal wind-field is known. This can be regarded as the second major forecasting problem, that the sign of the divergence cannot be estimated from observed data.

The only available alternative to the indirect measure of wind from the pressure field is to measure it directly by radio-pilot, and to plot streamlines from the observations so obtained. In practice, however, constructing adequate streamlines from scattered wind arrows is very difficult. All that can be obtained is a crude picture of the general wind direction. It is impossible to derive the vorticity or divergence from maps drawn in this way. Until some indirect way of estimating the wind is available, however, there is no better device than a close network of radio-wind stations. An extended effort to draw charts in this way has been made in the West Indies by the Institute of Tropical Meteorology of the University of Chicago. The charts cannot be looked upon as very successful.

It is suggested that the following research projects are those most important as far as the wind-distribution is concerned:-

(i) a theoretical examination of the equations of motion in low latitudes, to account for the air motion Particular attention should be paid to the effect of the semidiurnal pressure wave, to the possibility of steady motion and to the movement of air across the equator;

(ii) a practical investigation in some low latitude area of the relation between wind and pressure. It is particularly important to determine how close to the equator the gradient wind equation is valid.

2. Fronts and moving bad-weather belts.

The equatorial belt is obviously unfavourable for the existence of fronts. Temperatures and humidities are uniformly high over vast areas, and it is hard to see how any great contrast between airmasses can be maintained for long. There are also good dynamical reasons for thinking that fronts in the equatorial belt are unlikely to be intense or long-lived. Most tropical forecasters nevertheless talk about fronts, and they figure on most synoptic charts. Front-like weather disturbances occur in many equatorial or sub-equatorial areas. The best known example is the so-called West African "tornado", a fierce westward moving line-squall of 100 to 200 miles length resembling a cold-front in every particular, though plainly no such thing. It is important that the physics of discontinuity surfaces in low latitudes should be clarified, and that the nature of the observed "fronts" should be more fully examined. The main points requiring attention are these:-

(i) the slope of frontal surfaces. The equations of frontal equilibrium show that the slope depends on the sine of the latitude, which vanishes at the equator. For a front to have appreciable slope in very low latitudes, strongly accelerated winds are needed. The special conditions of the wind field that permit of sloping surfaces of discontinuity should be examined theoretically. Efforts should also be made to measure the slope of any fronts that can be traced into equatorial latitudes.

(ii) the possibilities of frontogenesis in the Doldrum trough. It is fashionable at present to talk of the axis of this trough as the "intertropical front", where the two trades are supposed to meet along a sloping frontal surface. It is difficult to see how such a "front" can in any way resemble the ideal fronts of middle latitudes. Detailed observations in the vicinity fail to reveal any sort of shear-plane between the two trade winds. A recent American committee on tropical synoptic models has suggested the substitution of "intertropical convergence zone" for the offending term, with the concurrence of J. Bjerknes.

(iii) the nature of the real front-like disturbance should be examined on the spot, preferably with the aid of upper air soundings.

The usage of the word "front" has become loose in all latitudes, especially within the tropics. It is considered that it should be rigorously confined to surfaces having the essential dynamical properties of a discontinuity, and that its use in other contexts should be discouraged. Its use to describe any line of bad weather simply hides the real character of the disturbance.

3. Stability and Convection. Precipitation.

Most rain in equatorial latitudes falls from cumuliform cloud, either as isolated showers or thunderstorms or from massed cumuli forming wide rain areas. It is clear that an understanding of stability in equatorial airmasses is essential. It is often assumed that all airmasses in these latitudes are damp to the tropopause and highly unstable. This is now known to be the exception, not the rule. The dry upper troposphere characteristic of the trade-wind belts often extends over the doldrums, and the air column is often conditionally stable even for prolonged heating inland. Potential instability is nearly always present, and general uplift over high ground or in areas of convergence usually produces heavy rain. Even in these cases, however, there are wide variations in the intensity of the rain without any apparent corresponding change in the surface distribution of wind, temperature and humidity, clearly indicating variations in stability. Systematic investigation of stability is urgently needed, particularly as regards the humidity of the upper troposphere.

Rain is often observed to fall from cumulus or stratocumulus whose tops lie well below the freezing level. It is now generally agreed among meteorologists working in low latitudes that the Bergeron-Findeison mechanism for the coagulation of water and ice particles is not the only mechanism by which raindrops can be formed. An investigation of rain formation in these latitudes is urgently needed.

4. Tropical revolving storms

Although they primarily affect tropical rather than equatorial latitudes, cyclones, hurricanes and typhoons form one of the major unsolved problems of meteorology. It is not known how or why they form, what guides their movement, what controls their form, deepening and decay, or why they have pronounced seasonal variations and areas of occurrence. Norwegian meteorologists, arguing from analogy with extra-tropical cyclones, have tried to establish that they develop as waves on the so-called intertropical front, and the idea that they involve the differing airmasses is widespread. These efforts have been highly speculative, and have been made in the almost complete absence of observations of the internal structure of the storms. The interior of the storms will have to be investigated aerologically and by direct aircraft reconnaissance, principally in the regions of cyclogenesis. Flights similar to those made recently by Major Wexler and Colonel F.B. Wood* into the interior of an Atlantic hurricane would repay good dividends.

* Bulletin American Meteorological Society 26. p.152 May 1945.

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20th November, 1945

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

CONVECTION IN THEORY AND PRACTICE

REMARKS ON MRP 242

by

Sir Charles Normand

AIR WEATHER SERVICE
TECHNICAL REFERENCE FILES

1. Summary. The statistical investigations in Part III are very valuable and full of interest. Whenever the claim is made, however, that they test the relative merits of the slice and parcel methods, the conclusions seem to me not to be valid.

It is to be hoped that statistical investigations on somewhat similar lines will be made in other parts of the world, particularly in the tropics, inland as well as near the coast; also that special flights will be organised to study the physics of convection and in particular to study the several points which statistical investigations leave unsettled. Both gliders and aeroplanes seem to be required in such investigations.

Regarding the theoretical exposition in Part II, the slice-method is excellent in its way, just as the parcel-method is, or the Bénard cell. Each adds to one's insight into the phenomena of atmospheric convection. There is no reason for, and every reason against, trying to impose a clear-cut demarcation between the parcel and slice methods. It is quite possible and healthy, to regard the slice method as one of the several possible and natural extensions of the parcel method.

In the paper the authors have made the mistake once or twice of stating that the parcel method underestimates, in certain cases, the amount of available energy. Also when testing the single parcel method, the authors have generally chosen as their sample parcel one that possesses extreme values, instead of selecting a representative parcel.

In this note mention is made of other extensions of the parcel method, e.g. methods that give values for the average parcel and methods that allow for admixture of a rising parcel with the environment, i.e. for "eviction", in Sir Napier Shaw's phraseology.

2. Single -parcel method. Both the application and the limitations of this method seem clear enough. It has one great use in the ready classification of stability. It shows up quickly those portions of a column of air which have latent instability. It gets one into the habit of looking at two curves simultaneously - the dry and wet bulb curves - and of keeping in mind that the atmospheric working substance has two components, the disturbing water vapour in addition to air.

In the tropics, looking at the dry and wet curves separately, one sees conditional and convective instability present nearly all the time, but a glance at the two curves together reveals the more important information, viz, whether and how much latent instability is present.

Unlike the steam engine in which each successive unit of the working substance goes through the same cycle, the atmosphere has differing parcels with varying initial and final states. So a single parcel in the atmosphere, unless carefully selected, is not, prima facie, likely to provide a quantitative answer to questions about average available energy or heights or amounts of cloud. The choice of a representative sample or the averaging or integration of the effect of a number of parcels may provide a quantitative answer. It is to be noted however, that in order to test the parcel theory the authors of M.R.P.242 have in general selected the surface parcel, i.e. one with extreme properties and usually not representative. When seeking quantitative results, few, if any, users of the single parcel method can forget that it applies rigidly only to a small parcel going up through the environment without loss of heat, without admixture, without growing so large that compensating down currents have to be allowed for. Now, differing parcels have differing initial properties and none will ascend without mixing to some extent with the surroundings. So, if one wants the single-parcel to provide a quantitative answer, one introduces a modification to suit the special circumstances. The major modifications are for the following types of circumstance:-

(a) Each parcel goes through it own cycle. Either a representative parcel has to be selected or an average has to be computed.

(b) Admixture (or what Sir Napier Shaw called Eviction) takes place and has to be allowed for.

(c) So many bubbles or parcels go up simultaneously that the compensating down current becomes significant. This is the circumstance, to meet which J.Bjerknes first introduced the slice-method.

3. Average Parcels. Occasionally any parcel is representative and gives the numerical answer. For example, when the air between the surface and cloud base is thoroughly mixed up, any parcel provides a fair quantitative estimate of the height of the cloud base. Even when conditions are not as favourable as this, it is probable that with practice one could select the representative parcel for an estimation of cloud height. This is really a matter for statistical examination.

In calculations of available energy it is rather remarkable to see how much attention has been paid to the parcel from the surface and how little to the column as a whole. It is about 40 years ago since Margules first showed us how to calculate the maximum energy available to the average parcel out of an unstable arrangement of parcels.

/To

To obtain Margules' results on T-ϕ grams, a third dimension, that of mass, has to be added in order to overcome the atmosphere's inconvenient habit of making each parcel pursue its own cycle.

Take the simple case of unstable dry air. If AB in fig. α represents on a T-ϕ gram an unstable column of dry air, the maximum energy will be realised if it overturns adiabatically and becomes CD. The parcel from A goes to C and gains its quote of energy A.C.B; B goes to D gaining energy B.D.A; the parcel in the centre remains where it is without gain of energy. The total energy is represented by the volume of a pyramid based on A.C.B.D; dividing by the total mass of the column, one derives the energy realisable by the average unit parcel between A and B to be equivalent to 1/6th of the area A.C.B.D, or $\frac{1}{3}$ of the area A.C.B. In deriving this result we are still assuming perfect adiabatic conditions, yet the answer is only $\frac{1}{3}$ that given by the surface parcel, which is the one ordinarily chosen by the author of M.R.P. 242.

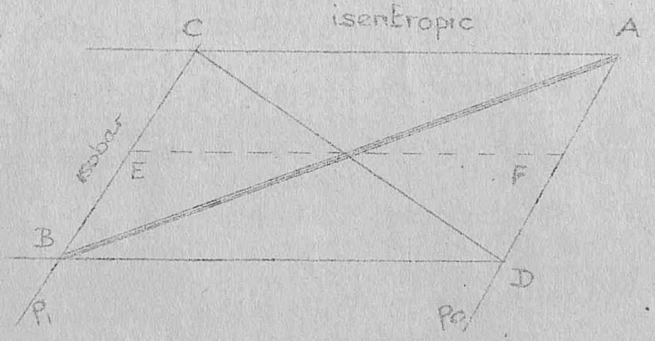

fig. α

The method of the average parcel can be applied to many cases. Here is a more general case. Suppose an initial unstable column such as A.B.C.D.E. If it rearranges itself adiabatically so that the parcels of lowest potential temperature go to the foot of the column and highest to the top and each in its potential order, then the maximum energy is, in theory, realised by the column, and the average energy per unit parcel is area BEDC X $(\lambda_1+\lambda_2/n)$

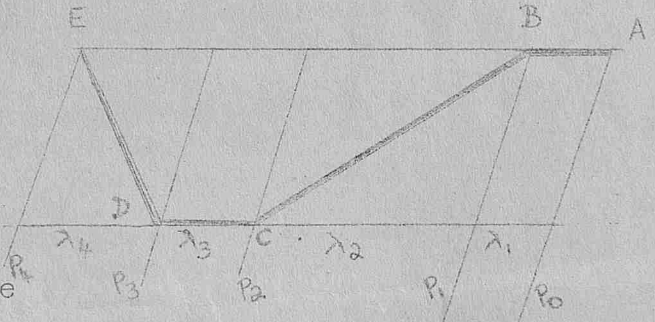

fig. β

where $n = \frac{1}{3}$ when $\lambda_3 = 0$, and rises to $\frac{1}{2}$ when λ_3 is large, and

where λ_1 is the ratio of the mass in A.B. to the mass of the whole column and so on, or $\lambda_1 = (p_0 - p_1)/(p_0 - p_4)$, $\lambda_2 = (p_1 - p_2)/(p_0 - p_4)$; $\lambda_3 = (p_2 - p_3)/(p_0 - p_4)$.
In the case where $\lambda_1 = 0$ and $\lambda_2 = \frac{1}{3}$, then the average parcel in the whole column has an available energy of about 1/8th of area B.E.D.C, i.e. only about 1/8th the energy of the surface parcel.

When air is saturated and remains so in descent, as in ascent, the same methods apply as above, except that saturated adiabats take the place of the dry adiabats in figs α and β, and the wet bulb curve constitutes the environment. In India in the monsoon a useful measure of latent energy is often furnished by regarding the wet bulb curve alone and treating it after the fashion of fig. β.

The main thing to note is that the average parcel usually provides an answer quite different from the surface parcel.

4. Admixture or Eviction. Another main factor affecting all quantitative results in problems of convection is the mixing of air. Sir Napier Shaw warned us often enough about "eviction", and it is difficult to imagine any upholder of the parcel method neglecting to make mental reservations on that score. No parcel is likely to rise in the atmosphere intact and unaltered in composition. But, what allowance is to be made for this? Consider again the simple case of dry unstable air. A simple calculation tells us that if a parcel during ascent keeps on growing so that its mass m $= ch^{\mu}$, then the lapse rate of the parcel is $(\gamma_d + \mu\gamma_0) / (1 + \mu)$. The interesting case is when $\mu = 1$, i.e. when the parcel increases steadily with height, i.e. m = c h, and when its lapse-rate is half-way between those of the original column and the dry adiabatic. With this constraint applied in fig α, the parcel rising from A finishes, not at C but at E, where CE = EB. The same result is reached in theory by the unit parcel rising intact and adiabatically to C and mixing at that level with another unit parcel from B. In either case, the final lapse-rate of the column as a whole is EF. A column in this final state has an adiabatic lapse-rate and this is a much more likely finish to the convective disturbance than the stable rearrangement, D.C.

The average available energy per unit parcel is in this case one-half that previously calculated, viz. 1/12th A.C.B.D or 1/6th A.C.B.

The contract between the two cases has thermodynamical significance. The former calculation regarded the atmosphere as a perfect thermodynamic machine, in which every parcel moved adiabatically to its own level, determined by its potential temperature; it told us the maximum available energy. This calculation is for a process that departs from the adiabatic and allows mixing to such a degree that the final lapse-rate is adiabatic. As convection arising from an unstable state in the free atmosphere is expected to continue until the lapse-rate is adiabatic or stable, this calculation seems to tell us the minimum energy realisable out of convection of dry air. The factor, one-half, connecting these two calculations applies to many other simple cases of dry air. Till a better is found, it is convenient to apply it generally. Thus, the average energy from the instability represented in fig. β becomes, when mixing is allowed for, very nearly $(\lambda_1/2 + \lambda_2/5) \times$ BEDC/5.

5. Balancing Vertical currents - moist or dry. Even a small bubble of air on rising must have its balancing down-current. So long as the bubble is small, the down-current is imagined to be distributed over a wide area and is therefore negligible. That is the single- parcel method. In the method of the average parcel, as illustrated in figs α and β, downward movements were accounted for as fully as the upward. The authors of M.R.P.242 however believe that computations by parcel method are not complete. At the foot of p.11 and on line 9 of p.12 they state in italics that the parcel method underestimates the amount of excess temperature in both dry and saturated convection. This is not so. An ascending parcel may be $2\,\delta T$ warmer than a descending parcel and only δT warmer then the static environment but against the static environment only one parcel, or one unit of mass, comes into the calculation. On the other hand, immediately the two parcels, one going up and one down, are contrasted two units of mass enter the calculation and the excess temperature of $2\,\delta T$ has to be shared by two units of mass.

Again, on p.14 line 19 and on p.36 line 9 same mistake is made, only it is available energy which the parcel method is here stated to underestimate. See in particular on p.36 where it is stated that "the parcel method will give only about 50 per cent of the true amount of energy." Again one is forced to repeat that the single parcel method does not underestimate. The double amount of energy found by the slice-method comes from the double unit of mass, one ascending and one descending.

6. Moist ascent, dry descent, slice-method.

When up-currents are saturated and down-currents dry and when each is substantial in area and volume, special difficulties arise in selecting representative parcels and in calculating average energies. J. Bjerknes (1938) threw light on this problem when he published his analysis, that has become known as the slice-method. Since then, the contrast between slice and parcel methods has been over-emphasised. The simplest interpretation of the slice-method is that given by the method and in the language of parcels. M.R.P.242 adopts this "simple derivation of the essentials" of the slice-method above fig.3 on p.10 when one parcel of mass M is moved down from B to C and another of mass M up from A to C. Similar arguments are used below fig.4, on p.11. Air from A goes up to C and other air from B down to E. The slice-method can be interpreted as the method of two unequal parcels in contrast to the one parcel in the single-parcel method.

The slice-method is a necessary extension or modification of the single-parcel method, but like all other methods has its limitations. They arise from the assumptions of (a) no sideways motion and (b) no descent within the cloud. There must be sideways motion at some levels, which may carry cloud sideways from the ascending current. It is known too that Cu-nimb and some other Cu clouds have descending, as well as ascending, air within them. The Kinematic models of Cu-nimb clouds vary from text-book to text-book. Some have the down-current in the middle, some at the back, some in front and some all around the core.

They all agree, at least, in showing a down current within the cloud. This upsets the argument about cloud amount in the slice-method, in which only up currents are assumed to have cloud. Nature herself never seems to be sure about how much down current should be dry and how much saturated.

7. Estimates of cloud tops. When the slice-method postulates no sideways motion, it seems necessary to postulate that each up current (and down current too) is of infinite height. This affects the argument on the top half of p.13. of M.R.P. 242, for where is the level of D by the time the cloud reaches it? Further the equations in the top half of p.13 seem to apply no more particularly to slices than to parcels. They are independent of W/W' and of M/M'. They apply just as well to a parcel starting a little way below D and rising a little above D. The results all follow from the constraints here postulated. These constraints are that $\partial w/\partial z$ remains negative above D and W equals zero some height above D. One may similarly, in the parcel method introduce constraints. It is clear, for instance, that the AA' of fig.2 of M.R.P. 242 is not a representative parcel, because the wet-bulb curve A' Tw falls away from the saturated adiabat, A'B. The representative sample therefore has its adiabat to the left of A'BE; that makes the representative E lower than is shown in fig.2. Again admixture or eviction makes the rising air follow a course still nearer that of the environment. Add the effect of the wind shear above D. It is just as logical to conclude that a cloud top nearer D than E follows from a parcel method as readily as it does from the slice-method. Anyhow, a contrast of the two methods by this means seems rather futile.

8. The Statistical Characterisitics of Part III, M.R.P.242. This part is full of most interesting and useful information, and the only major point on which one can disagree is the claim that the relative merits of the slice and parcel theories are tested herein. The unmodified single-parcel method works quite well, it is clear, for occasions for which it is suited (see § 17 pp. 20-24). Nearly all the other applications of parcel method are made with unrepresentative samples. There's an exception in § 19 at top of p.27, where a representative value is selected.

The Table XXVII, p.33, displays a very suggestive correlation between equivalent cloud amount and degree of instability and one wonders whether the scatter would diminish and the correlation be improved, if representative rather than surface parcels were selected or if the energy of the average parcel had been calculated.

About the general conclusions on p.36 I'm not clear. The authors state that the ascent of saturated air as a whole is slow, but that vertical gusts of 6 m/s may be fairly frequent. American gliders say that even in mild Cu-nimb clouds they rise 3,000 ft. in a minute; they have ascended 20,000 ft. in 5 minutes. In India Mitchells and Dakotas are reported to have been sucked up 10,000 ft. while the pilots were trying to fly level. Considering the speeds of aeroplanes, and of sailplanes too, can one say these sudden ascents are the results of gusts within the cloud? If so, it would seem that a single gust occupies a large proportion of the cloud's volume.

The vertical velocities calculated in Table XXX from the single parcel theory are again those from extreme samples, not from representative or average parcels; nor is allowance made for mixing. It does not seem quite correct to say that these extreme velocities (see foot of p.35) "represent the mean values for the entire mass of saturated air". They do not do so.

There are three minor points - two question and a suggestion. The questions are - in Table XVII can the last column have a zero entry against 900 - 940 mb., when the second last column has the entry 2; and why under observed cloud top are no entries made below 440 mb, when Table XXXVIII shows many cloud tops reaching beyond 440 mb? It is suggested that in Table VII p.20 the interchange of Cb and Cu in both columns and rows would improve it as a correlation table.

9. Some further investigations. In § 24 and Table XXVIII an examination is made of Burgeron's theory of ice-crystals and super-cooled water being necessary for the release of precipitation. In England however all deep clouds necessarily reach beyond the freezing level, and the results on this point are therefore not conclusive. A similar examination in the tropics is likely to provide a better test to the theory. The most hopeful area that I know for the test is in the field of the monsoon current from the Arabian Sea over India. Within this current, when the monsoon is properly established, convective showers are plentiful, but thunder showers are absent. The working hypothesis there might be - is ice necessary for thunder, but not for convective rain?

The high lapse rates found within the clouds, Table XXIX p.35 are remarkable, particularly when it is remembered that the pilots may have been automatically selecting against the higher lapse-rates. Is this connected with the old question of clouds being often cooler than the surrounding air? Kochanski (Inst. Geophys and Met. Lwow, Vol.8, p.311, 1936) had a long paper on Thermal currents and Cu clouds, with a statistical examination of p.b. ascents and some aeroplane soundings. He also reported the high lapse-rates within clouds and stated that the interior of Cu cloud is always colder than that of the air outside at the same levels. More comparisons of air within and outside clouds are desirable.

Both dry and wet-bulb temperatures are required in the comparison. Where a cloud mixes with surrounding dry air its temperature should tend towards equilibrium with the wet-bulb of the surrounding air. Or, is it mainly an effort to equalise virtual temperature? Anyhow, we require a comparison of wet-bulbs as well as other temperatures inside and outside clouds. It is obvious that we have still much to learn about convection.

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MRP 270

REPORT ON A QUICK RESPONSE THERMOCOUPLE PSYCHROMETER FOR USE ON CAPTIVE BALLOONS.

by

W. C. SWINBANK.

~~Publd in Journal of Sci Instruments,~~
~~Vol. 18, No. 7, July 1941,~~
~~pp. 119-125~~

G.102178.

An Automatic Recording Thermocouple Psychrometer for use on Captive Balloons.

1. Introduction.

An instrument for use on captive balloons which records either the temperature and humidity behaviour of the air at a fixed point or the vertical profile of temperature and humidity in the atmosphere up to heights of 6,000 feet has been developed at R. and D.E., Cardington, in conjunction with the Meteorological Office.

The heat capacity of the thermometric elements is so small that the instrument shows 90 per cent response to temperature change in two or three seconds. It is therefore particularly suitable for the investigation of the fine structure of the atmosphere in connection with turbulence and allied problems, and it has also provided striking records of the detailed structure of subsidence inversions. Examples of both types of record are given later in this report.

2. The Psychrometer.

The psychrometer, which is designed on the Assman principle, is shown diagrammatically at Figure I. The thermometric elements consist of two sets of thermocouples, one to serve as the dry and the other as the wet bulb thermometer. The warm junctions of the thermocouples are situated, as shown, in a cylindrical plastic tube, one on either side of a plastic strip that divides the tube longitudinally. The reference, or cold, junction of each thermocouple is maintained at a constant temperature of 32°F by immersion in a test tube of liquid paraffin, the test tube itself being immersed in a mixture of crushed ice and water packed in a quart Dewar flask. It was found during the early development of the instrument that considerable temperature gradients could exist in the Dewar flask, due, presumably, to the anomalous density property of water at temperatures just above freezing. This fault has been obviated by the introduction of the heat distributor shown in the diagram. It consists of a series of copper fins of large area whose conductive property reduces the temperature differences in the flask.

Each set of thermocouples consists of four copper-constantan thermocouples connected in series; the total E.M.F. from each set is 0.09 millivolt per °F difference between warm and cold junction. The instrument has been designed primarily to meet the requirement of quick response, and this is most obviously to be achieved by using very fine wires in the construction of the thermocouples. But it is also desirable that the response times of both dry and wet bulb thermometer should be, as far as possible, the same. This requirement imposes a lower limit on the response time that can be attained because of the thermal inertia of the wick surrounding the wet bulb which cannot be reduced below a certain minimum necessary for the efficient working of the wet bulb thermometer. It has been found by experiment that the optimum thickness of wire that meets the double requirement of quickness of response and equality of response for dry and wet bulb is about 38 S.W.G.

The thermocouple junctions are made with Wood's metal, and the lead wires are supported in the tube by attachment at intervals to a narrow strip of cellulose acetate. The wires are cotton covered and shellaced; the shellac improves the insulation and prevents the paraffin at the cold junction from creeping along the cotton. The wet bulb junction is covered with thin muslin which is fed by a wick leading to a reservoir containing distilled water.

Aspiration of the thermocouples is provided by a fan at the head of the plastic tube which draws a current of air past the junctions at a speed of about 35 feet per second. The fan is driven by a 12 volt D.C. motor.

The complete psychrometer unit is assembled in a frame of tubular steel equipped for attachment to a balloon cable. The photograph at Figure II shows the apparatus mounted on a balloon cable ready for use.

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3. Method of Measurement.

The currents generated by the thermocouples are conducted separately through a four core cable sheathed in the high tensile steel wire that forms the balloon cable to the recording equipment on the ground. At the ground the four cores pass through the shaft of the balloon cable storage drum to four copper discs, insulated from each other and running in mercury, thus allowing records to be made whilst the balloon is being raised or hauled in.

Two methods of recording are used, one visual, the other automatic. The first, which measures the E.M.F. by means of a null-deflection potentiometer, was satisfactory with the slow response psychrometer that has been used previously; moreover, it has the advantage that the resistance of the cable leads is immaterial. But this method is difficult and unsatisfactory for the often rapidly fluctuating E.M.F's produced by the new quick response instrument.

In the second method the E.M.F. from each thermocouple is applied to a Tinsley Mirror Galvanometer. The light spots from the two galvanometer mirrors are focussed by means of a cylindrical lens onto a sensitized bromide paper carried on a rotating drum. The drum is driven at a speed of one rotation per hour by a synchronous motor. Thus an automatic photographic record is made of the varying wet and dry bulb temperatures of the air encountered by the psychrometer. The whole of this recording apparatus is assembled in a dark room.

The assembled recorder is shown as part of the photograph at Figure III.

4. Method of Calibrating the Recorder.

The recorder is calibrated by impressing on it a known E.M.F. to simulate the E.M.F. produced by a given temperature difference between the warm and cold junctions of the thermocouples. The circuit diagram of the apparatus used in the calibration is shown at Figure IV. A step by step potential divider is used to give E.M.F's equivalent to those produced by 10°F intervals of temperature difference between the warm and the cold junction. In the calibration circuit is a 500 ohm variable resistance which is set to represent the total resistance of the thermocouple and the cable leads to the recorder.

5. Lag of the Instrument.

Figure V shows traces recorded on a rapidly rotating drum by the dry and wet bulb thermocouples when subjected to sudden changes in temperature and wet bulb temperature respectively. It will be seen that in both cases 50 per cent of a given temperature change is recorded in little over a second, whilst 90 per cent of the change is recorded in three or four seconds.

6. Records obtained with the Instrument.

The instrument has proved capable of providing records of very fine detail of the temperature and humidity structure of the atmosphere. Figure VI, which is typical of many records obtained, shows fluctuations of dry and wet bulb temperatures of sometimes more than 2°F in periods of about 10 seconds. These fluctuations are even more striking when it is remembered that the actual variations of air temperature and wet bulb necessary to produce fluctuations of this period on an instrument with a lag of about 3 seconds are about twice as great as those recorded, i.e. in this case the corresponding fluctuations of air temperature are about 4°F.

Figure VII shows a temperature and wet bulb temperature profile through a subsidence inversion. It will be seen that there is a temperature inversion of about 10°F in 100 feet. The comparative constancy of the wet bulb temperature through the inversion means that the dew point profile is almost a mirror image of the dry bulb profile. Several such records have been obtained at Cardington. A more recent sounding showed an even more pronounced

/inversion

[illegible] sharply [illegible] 9/1/46

inversion of about 12°F in 60 feet. So far as is known this sharpness of inversion, which shows a close approach to conditions associated with idealised surfaces, has not previously been recorded.

7. Accuracy of Recording.

Comparisons with a standard Assman psychrometer have shown that the instrument gives readings accurate to within ±0.2°F.

The present method of recording permits a discrimination of between 0.1 and 0.2°F.

8. Further Development.

(a) It is proposed to replace the single walled aspirating plastic tube with a double walled tube to serve as a radiation shield. A louvred housing has been used in the past, but for various reasons it has proved unsatisfactory.

(b) In order to reduce the weight of the instrument, and thus to enable greater heights to be sounded, it is proposed to replace the tubular steel frame with dural. Also, a pint flask will be substituted for the present quart Dewar flask.

(c) So that the response of the instrument can be still further quickened, the rate of aspiration is to be increased to 60 feet per second.

(d) The response time of the recording apparatus itself, with the circuit resistance at present used, is of the same order as that of the thermocouples. Whilst this is not serious with the present lag of the psychrometer, it will become so if the lag is still further reduced. It is therefore proposed to experiment with a cathode ray oscillograph method of recording.

Meteorological Office,
Air Ministry,
9th January, 1946.

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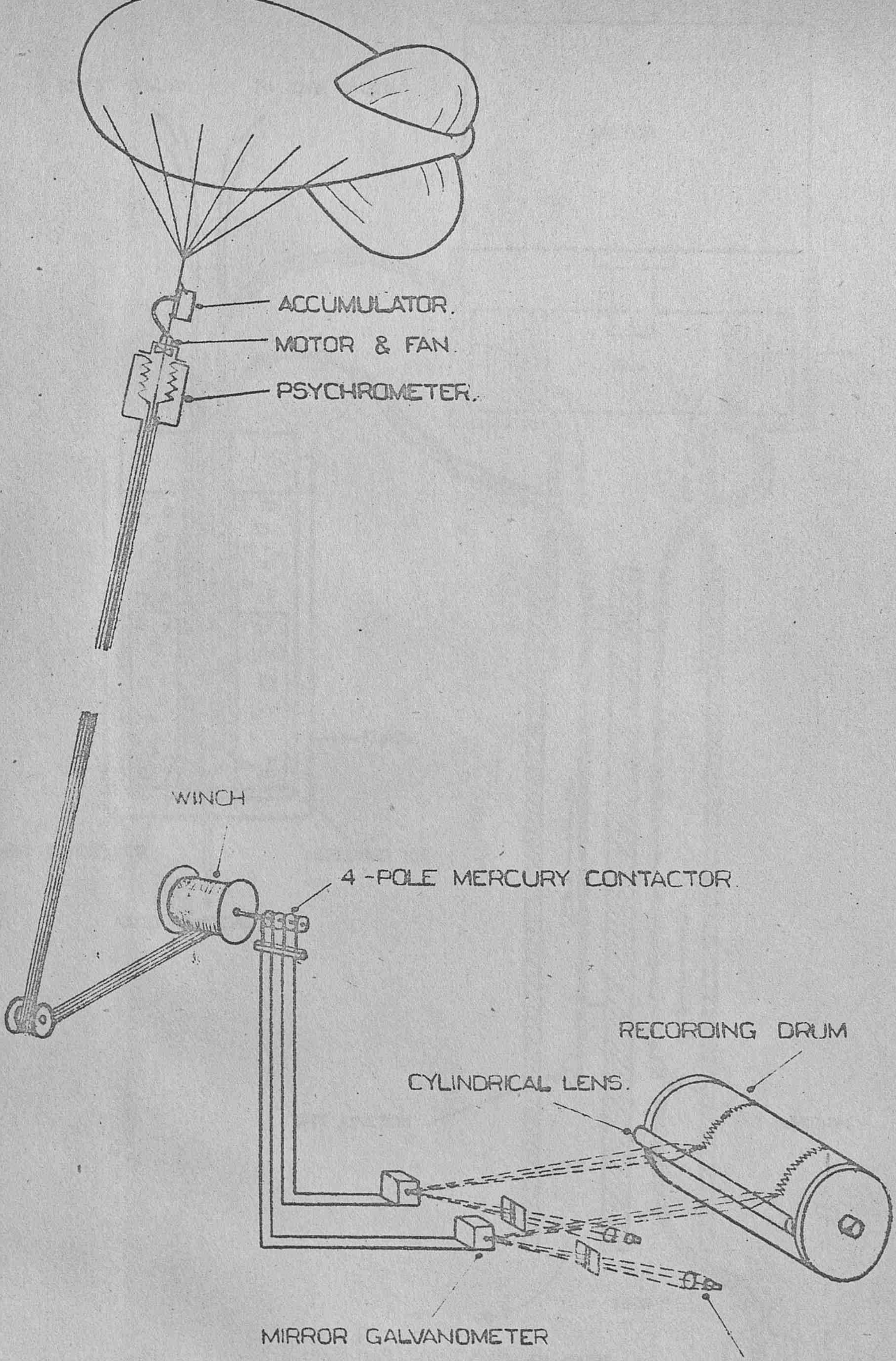
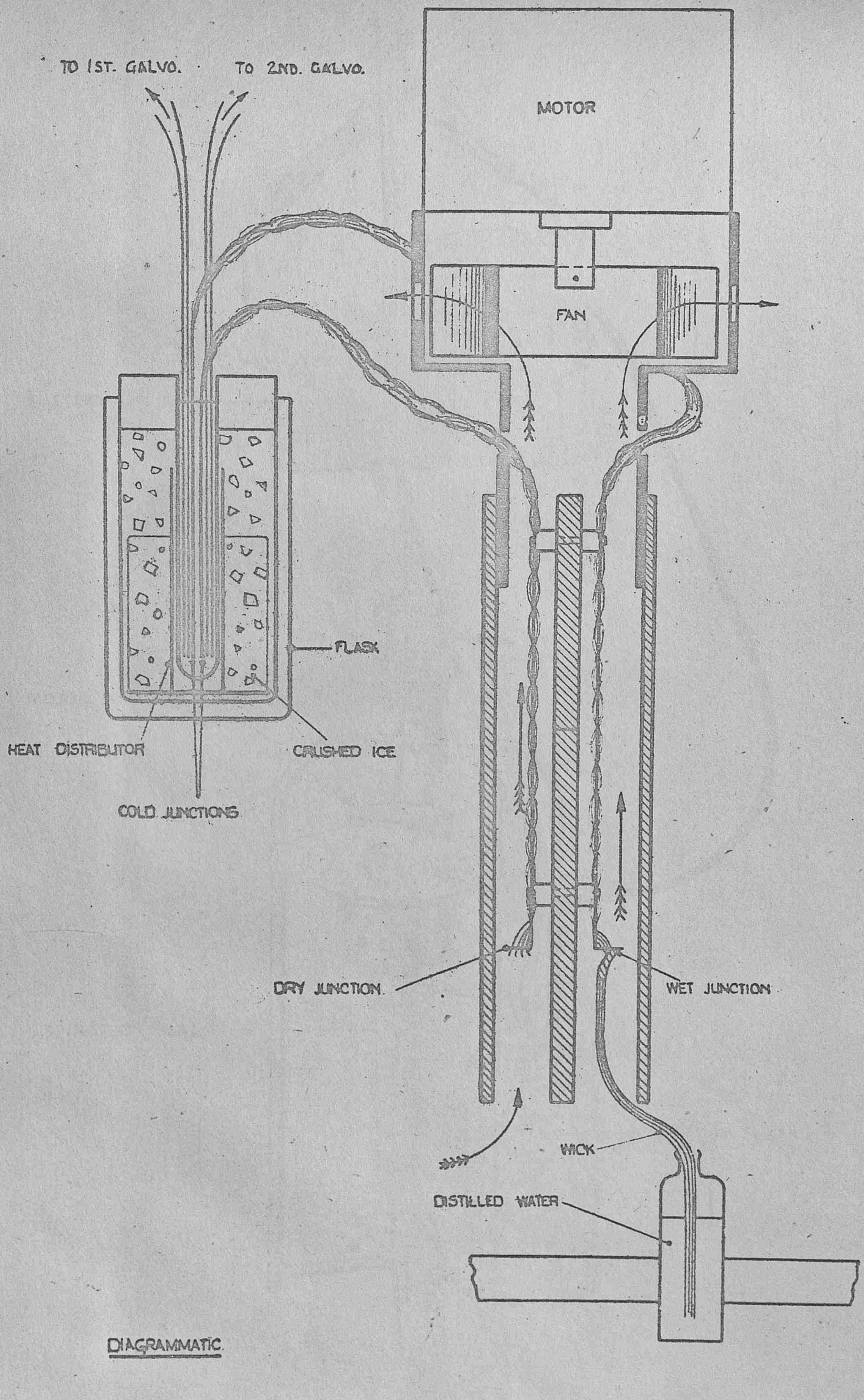

ACCUMULATOR.
MOTOR & FAN.
PSYCHROMETER.
WINCH
4-POLE MERCURY CONTACTOR.
RECORDING DRUM
CYLINDRICAL LENS.
MIRROR GALVANOMETER
LIGHT SOURCE.

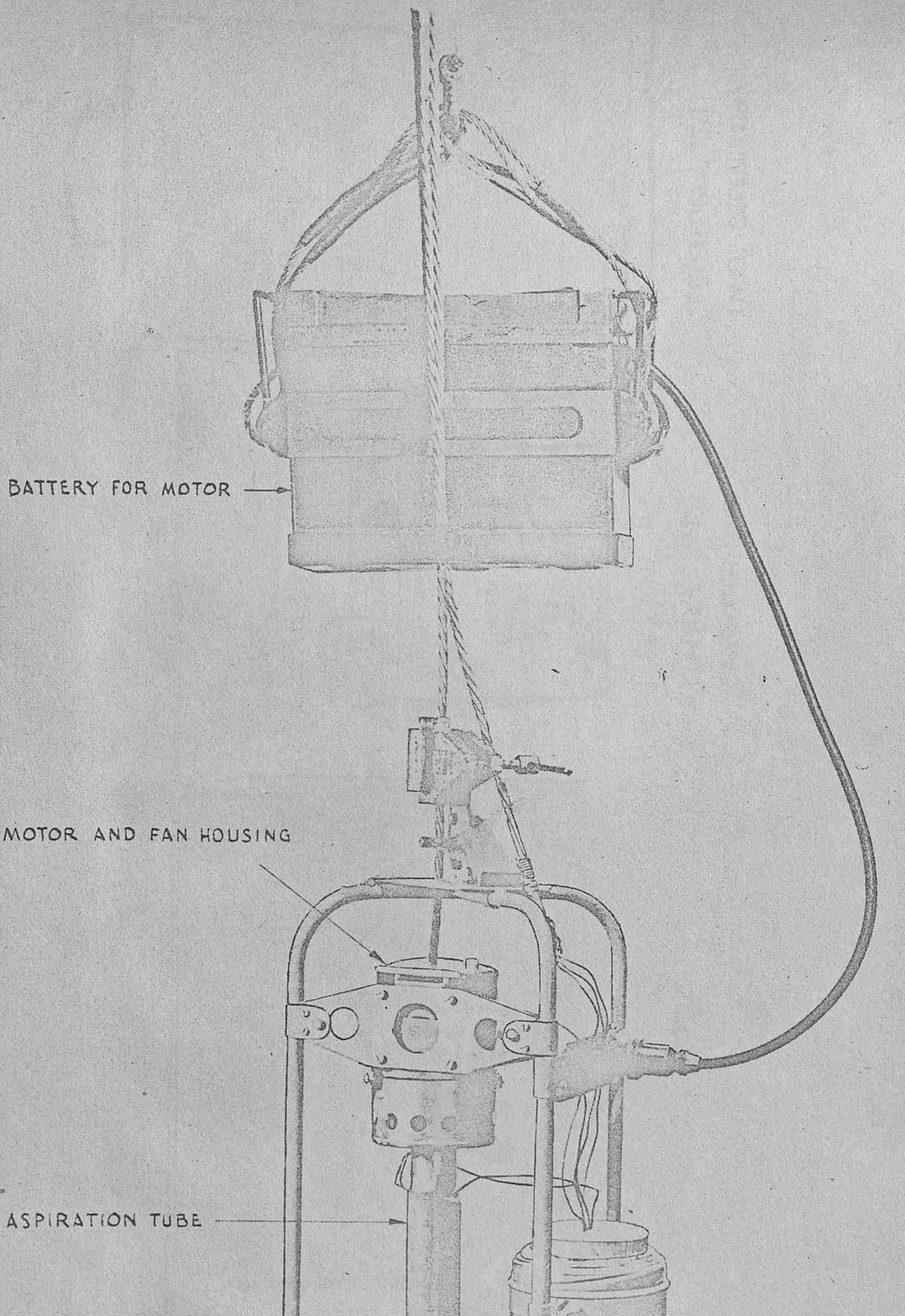
BALLOON PSYCHROMETER & RECORDING APPARATUS.

SCHEMATIC


TO 1ST. GALVO.
TO 2ND. GALVO.
MOTOR
FAN
FLASK
HEAT DISTRIBUTOR
CRUSHED ICE
COLD JUNCTIONS
DRY JUNCTION.
WET JUNCTION
WICK
DISTILLED WATER

DIAGRAMMATIC.

THERMOCOUPLE PSYCHROMETER FIG. I


BATTERY FOR MOTOR
MOTOR AND FAN HOUSING
ASPIRATION TUBE
THERMOS FLASK

PSYCHROMETER IN POSITION ON BALLOON CABLE

FIG II

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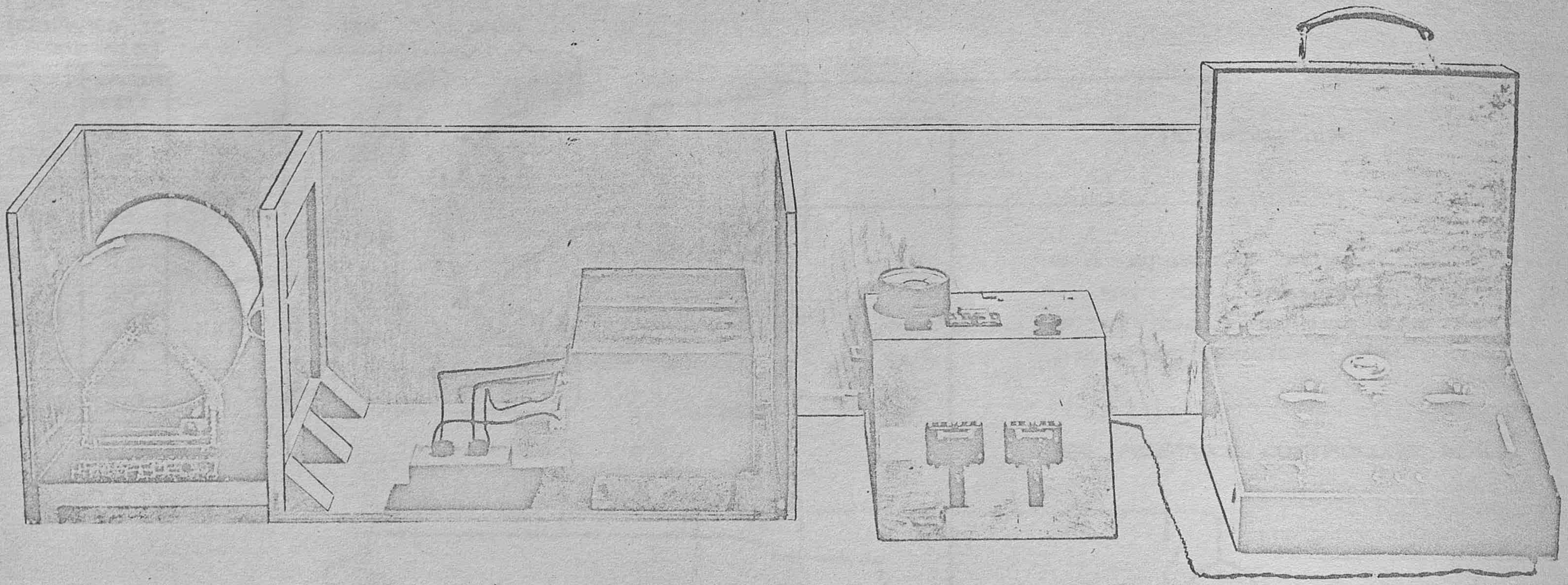
RECORDING DRUM
GALVANOMETERS
CONTROL AND CALIBRATOR UNIT
POTENTIOMETER (NULL DEFLECTION)

RECORDING UNIT FOR BALLOON PSYCHROMETER

FIG. III

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SCALE OF E M Fs & EQUIVALENT TEMPS AT CONTACT POINTS ON THE 1. OHM. POTENTIAL DIVIDER (COMPARED TO CONTACT POINT Nº 5.)

CONTACT POINT	MILLIVOLTS DROP	EQUIVALENT TEMP °F
1	-1·43	15°
2	-1·01	20°
3	-·60	25°
4	-·18	30°
5	0	32°
6	+·24	35°
7	+·67	40°
8	1·10	45°
9	1·54	50°
10	2·43	60°
11	3-32	70°

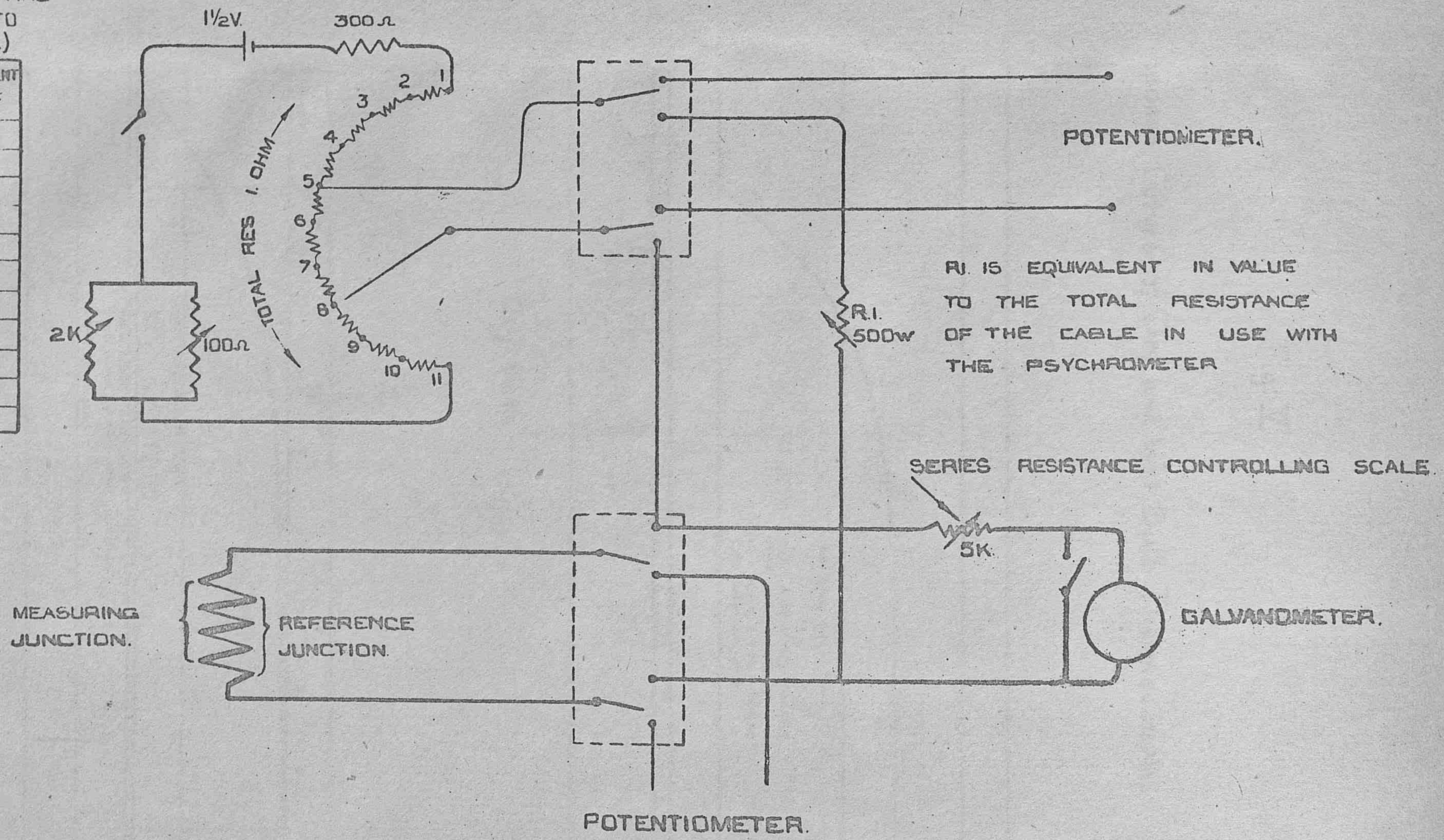
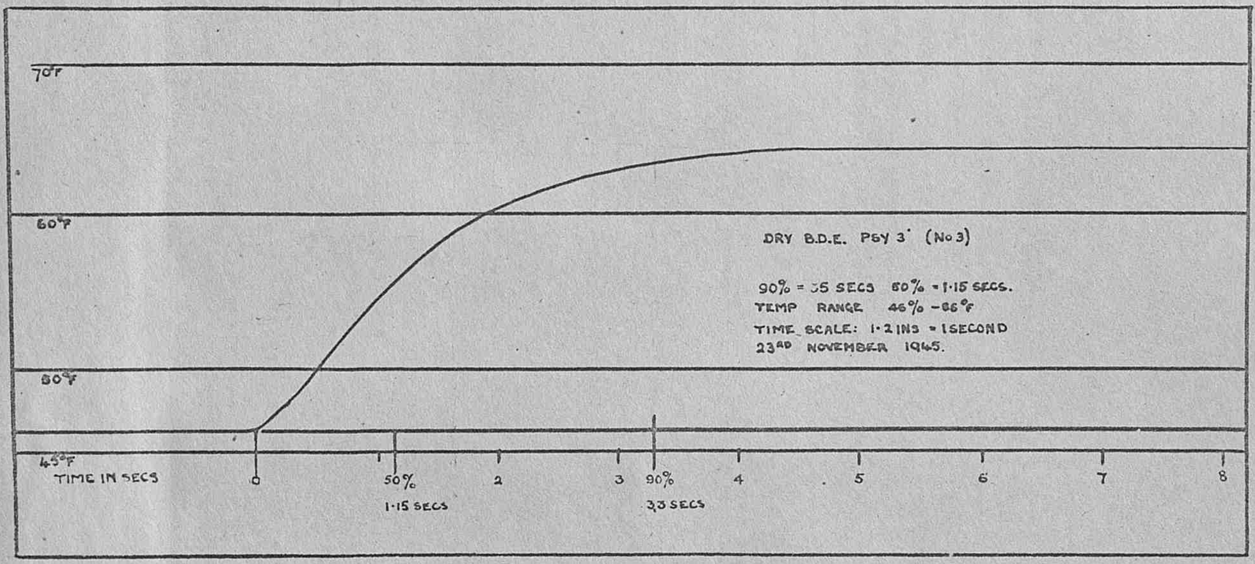

1½V.
300Ω
1
2
3
4
5
6
7
8
9
10
11
TOTAL RES 1. OHM
2K
100Ω
POTENTIOMETER.
R.1.
500W
RI. IS EQUIVALENT IN VALUE TO THE TOTAL RESISTANCE OF THE CABLE IN USE WITH THE PSYCHROMETER
SERIES RESISTANCE CONTROLLING SCALE.
5K.
GALVANOMETER.
MEASURING JUNCTION.
REFERENCE JUNCTION.
POTENTIOMETER.

FIG. IV

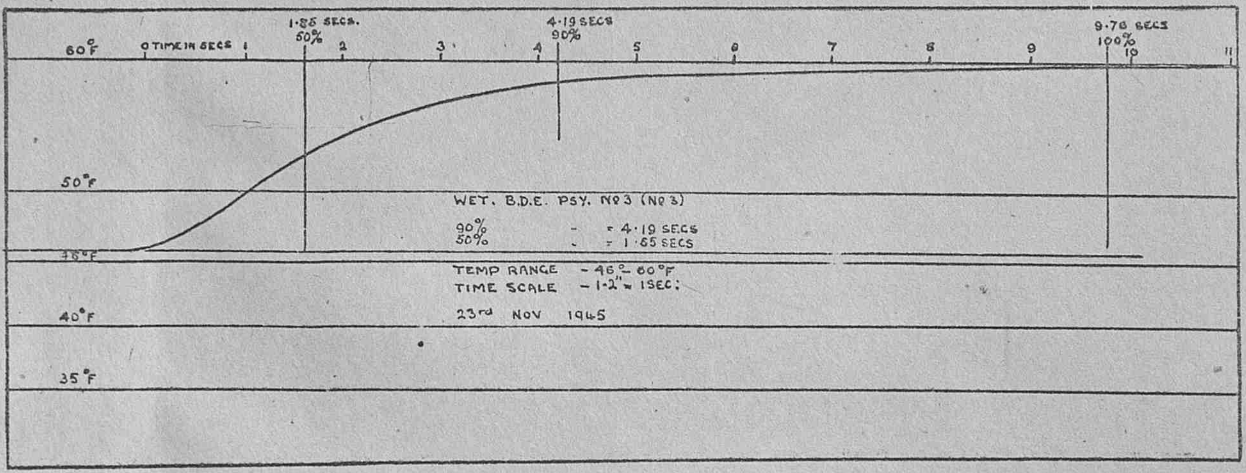
CALIBRATOR & RECORDING CIRCUIT. (ESSENTAIL DETAILS, ONE CHANNEL.)

Fig: V

Response Curves of Dry and Wet Bulb Thermocouples



Dry Bulb.



Wet Bulb.

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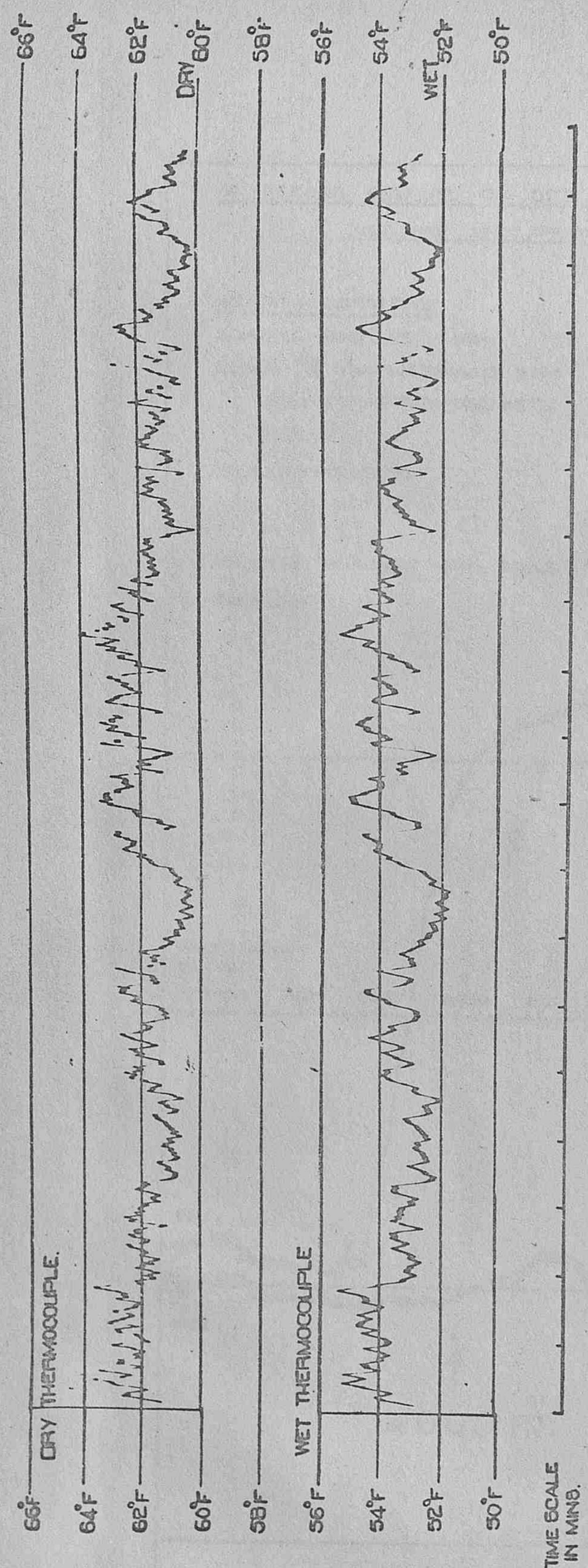


A DETAILED RECORD OF DRY AND WET THERMOCOUPLES TAKEN UNDER VARIABLE SUNNY CONDITIONS.

DATE: 4 · 6 · 45
TIME: 1015 HRS TO 1034 HRS. GMT.
WEATHER: $\frac{4}{10}$THS $\frac{5}{10}$THS FAIR WEATHER CUMULUS, SURFACE WINDSPEED, 15 MPH, GUSTY.

INSTRUMENT: B.D.E. PSYCHROMETER WITH DRY AND WET THERMOCOUPLES OF 36 SWG. COPPER/40SWG EUREKA, WELL ASPIRATED. THE INSTRUMENT DURING THE EXPERIMENT WAS STANDING ON AN IRON GRID ON A TRESTLE 4FT. 6INS ABOVE GROUND.

TEMPERATURE & HUMIDITY RECORD NEAR EARTH'S SURFACE

FIG VI

A RECORD OBTAINED ON OCT 1ST 1945 AT
1010 HRS G.M.T. APPROX.

WEATHER CONDITIONS:
SURFACE WIND: NE 1 MPH
CLOUD: $\frac{10}{10}$ STRATOCUMULLIS BASE
3500 FT APPROX TENDING TO
BREAK.
VISIBILITY: 2 MILES

HEIGHTS INDICATED ARE CABLE LENGTHS
PAID OUT.

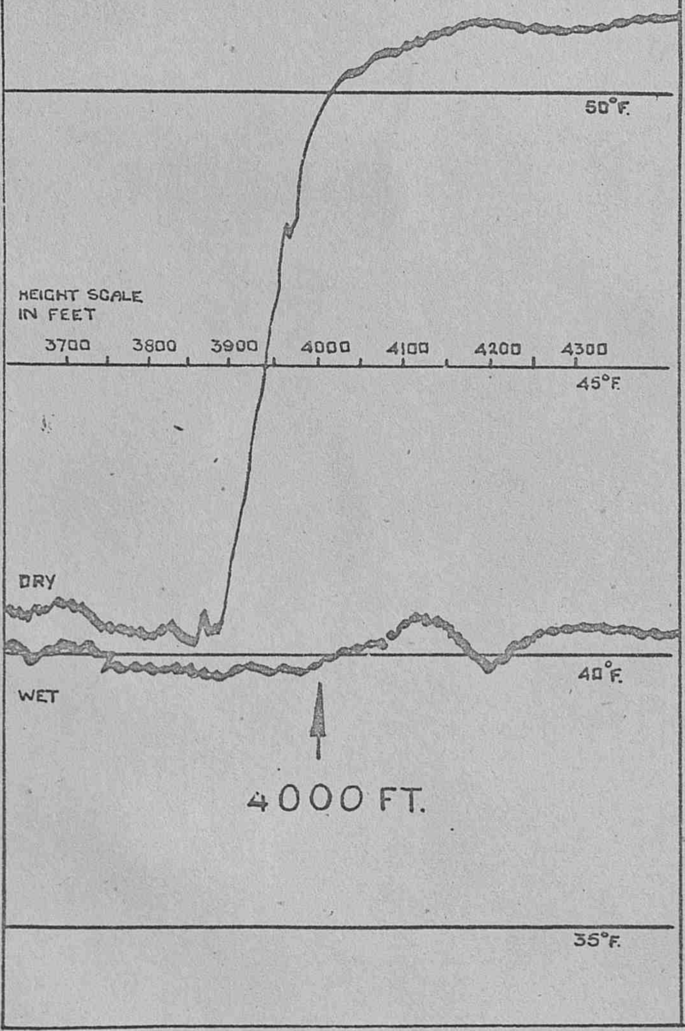


FIG VII

(Pt. I. follows)

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE.

The occurrence of exceptionally dry air in the trophosphere.

Part II. Observations with balloon psychrometer

by

W.C. Swinbank

The attached records of marked dryness above subsidence inversions have been made at Cardington with the balloon thermocouple psychrometer described in M.R.P. 270.

The saturation points have been calculated from the Assman psychrometric formula using the recorded values of wet and dry bulb temperatures. For saturation temperatures above 32°F the dew point has been plotted, and for temperatures below 32°F the hoar frost point.

At the time of the ascent shown in fig.4, a large anticyclone covered the British Isles, the maximum pressure being 1038 mbs. 60 miles to the west of Cardington. On the occasion of the ascent illustrated in fig. 5 an anticyclone stretched from N.W. to S.E. across the British Isles. The highest pressure was about 1032 mbs. over Northern Ireland and there were indications of a second centre with a pressure of about 1030.5 mbs. near Cardington.

G.102794.

MRP 272

14th February, 1946

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

The occurrence of exceptionally dry air in the troposphere

Part I. Observations with frost point hygrometer

by

A.W. Brewer

During the past three years, as part of the work of the development of the dew point or frost point hygrometer, aircraft ascents have been made under various conditions and to various heights up to 41,500 feet.

The frost point hygrometer is especially suitable for measuring low humidities, since the frost points has an approximately logarithmic relation to the vapour pressure and there is no contraction of the scale at low relative humidities or low vapour pressures.

The extreme variability of the relative humidity of the upper air was obvious from the results and this note is submitted to present a selection of the ascents.

The ascents are shown graphically in Figs. 1,2, and 3 and explanatory comments are included in the figures. The corresponding relative humidity is shown along each dew or frost point observation.

Ascent 1. 17.1.46. Ascent to 18,300 ft. (500 mbs.) showing very dry subsided polar continental air and overrunning moist air. The relative humidity at 6,400 ft. (800 mbs.) is the lowest we have recorded in the troposphere.

Ascent 2. 6.3.45. Ascent to 20,500 ft. showing subsiding tropical air with the lower limit of subsidence at 7600 ft. rather higher than usual. Two separate types of air may be involved, with a separation at about 670 mbs.

Ascent 3. 7.2.43. Ascent to 36,000 ft. showing subsiding polar air showing dry air at all levels. The discontinuity of humidity between the friction layer and the subsided upper air is being removed by penetrative convection which is reaching 8300 ft. (750 mbs.). Note the low relative humidity of the stratosphere.

G.102794

FIG 1

Ascent I. 17.1.46 (1400 GMT)

Ascent to 18,300 ft. (500mbs) showing very dry subsided polar continental air and overrunning moist air. The relative humidity at 6,400 ft (800 mbs) is the lowest we have recorded in the troposphere

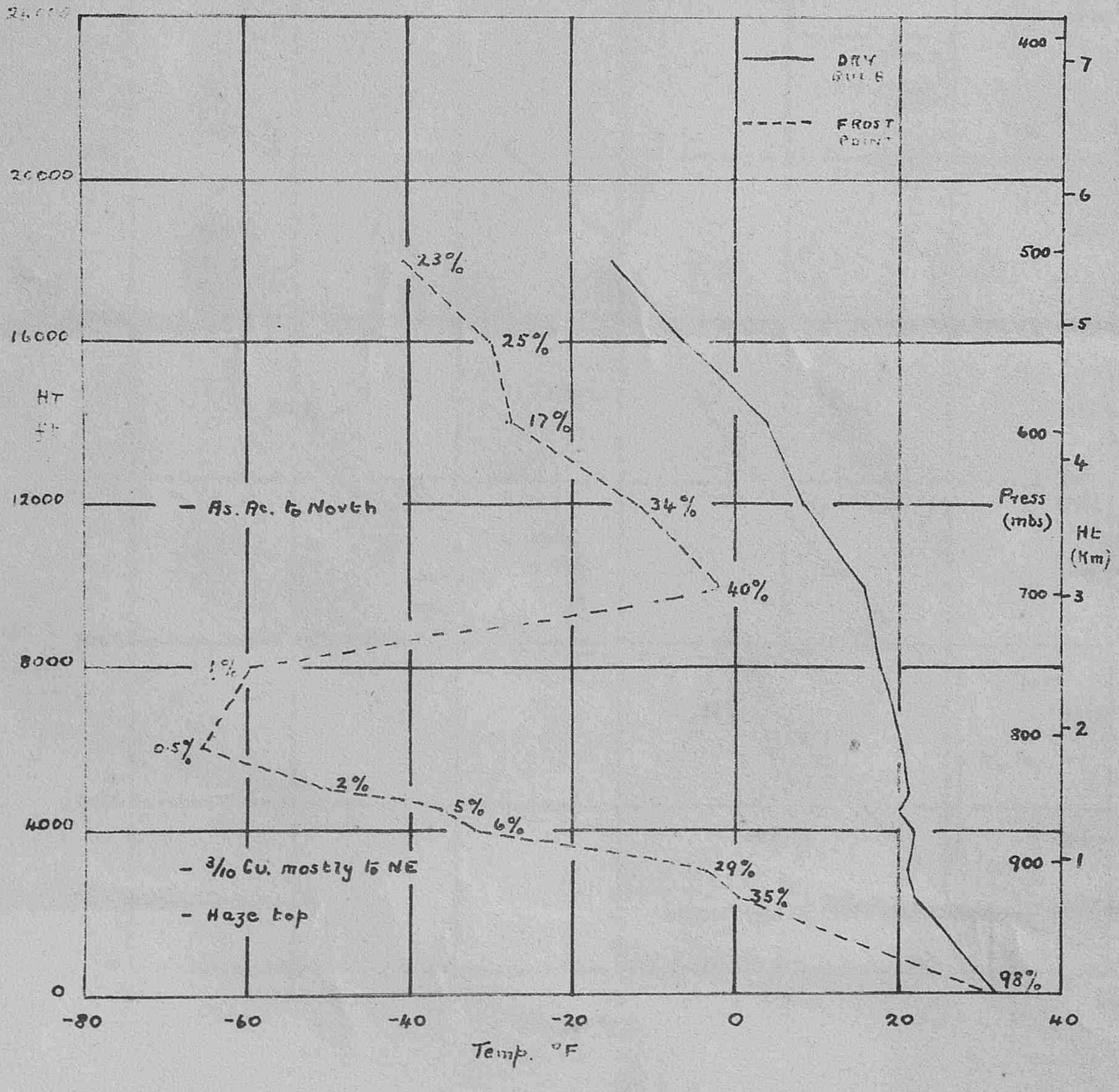


FIG 2

ASCENT 2. 6.3.45 (1500 GMT)

Ascent to 20,500 ft showing subsiding tropical air with the lower limit of subsidence at 7600 ft rather higher than usual. Two separate types of air may be involved, with a separation at about 670 mbs

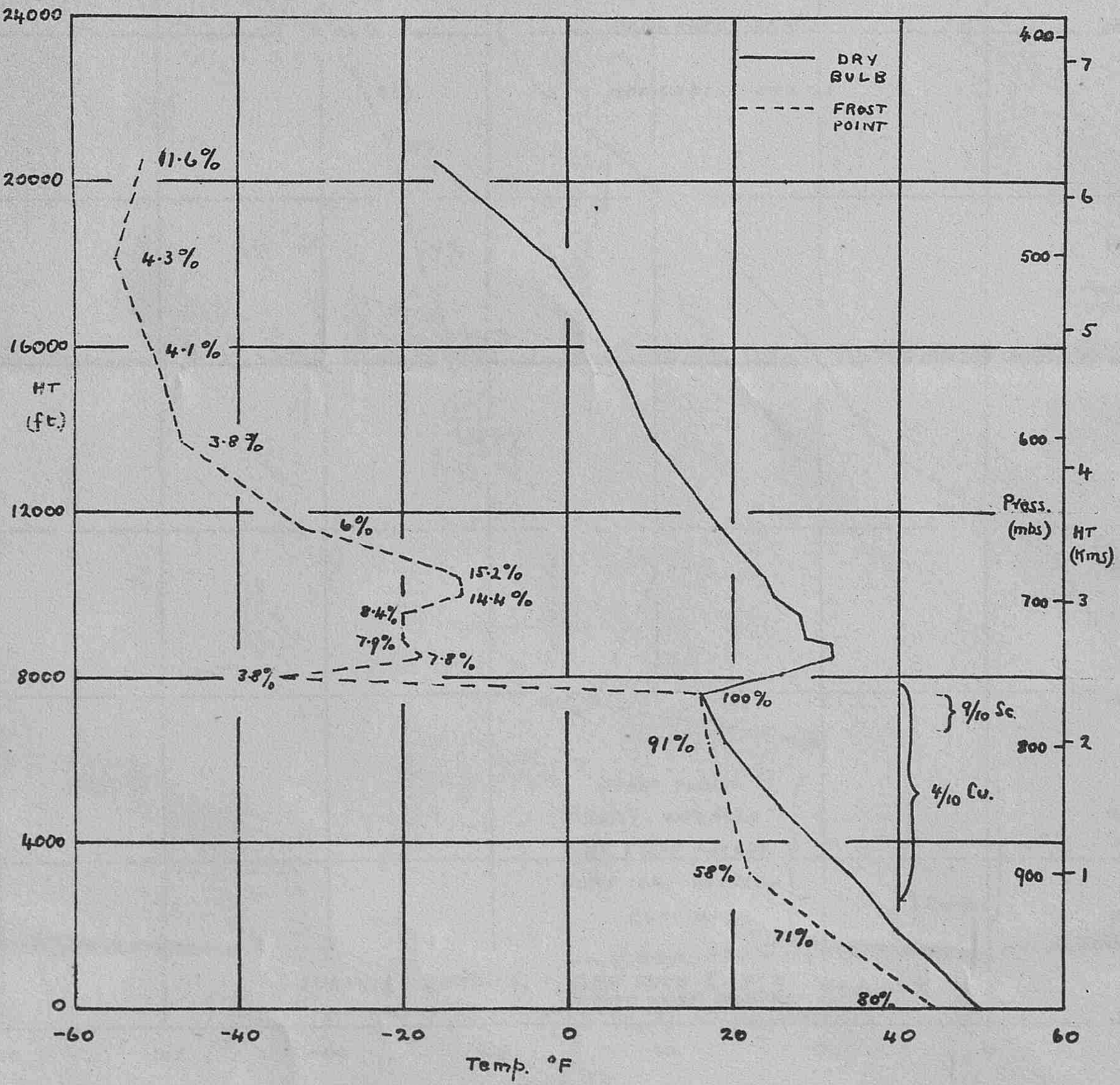


FIG 3

ASCENT 3. 7.2.43 (1300 GMT)

Ascent to 36,000 ft showing subsiding polar air showing dry air at all levels. The discontinuity of humidity between the friction layer & the subsided upper air is being removed by penetrative convection which is reaching 8300 ft (750 mbs). Note the low relative humidity of the stratosphere

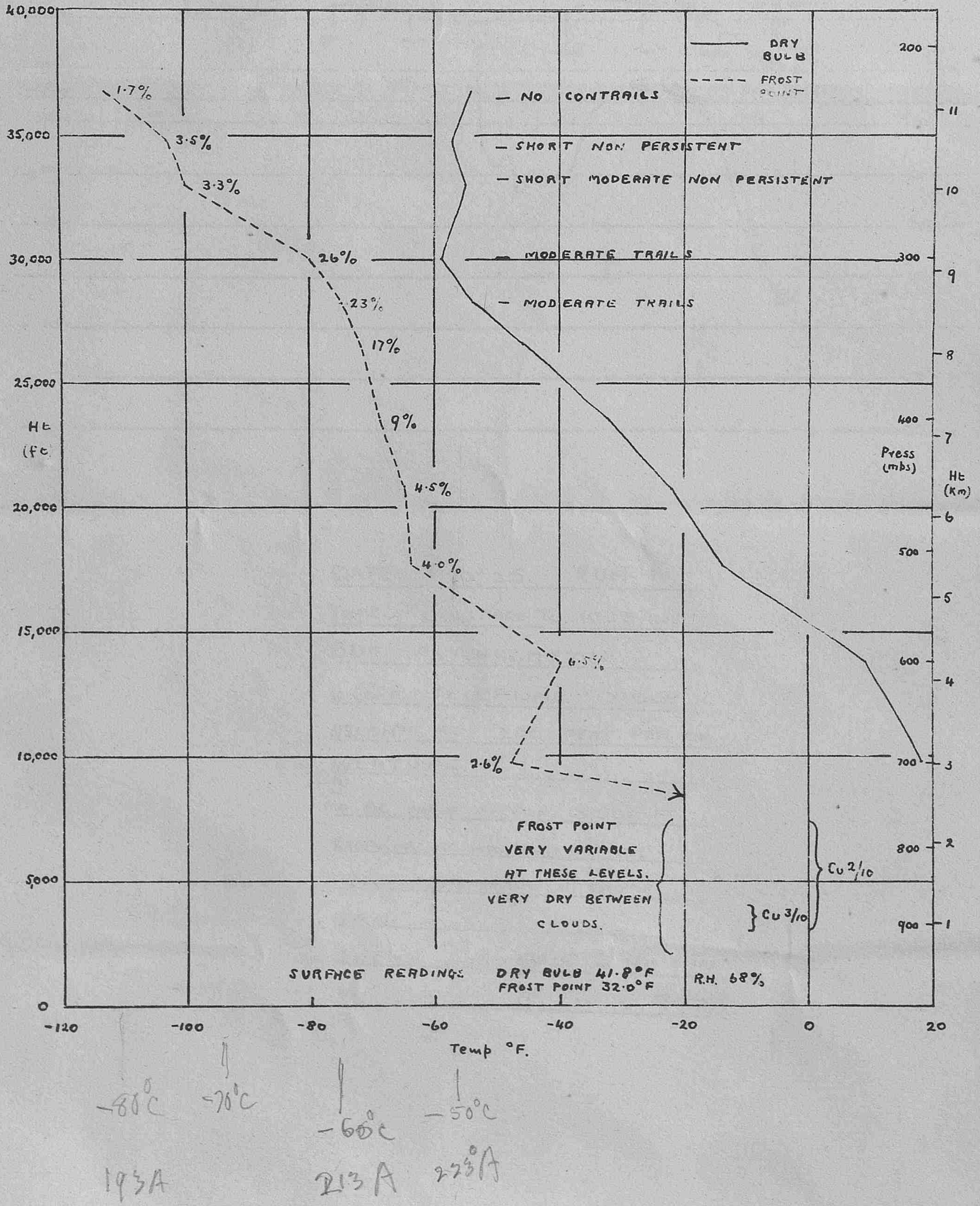
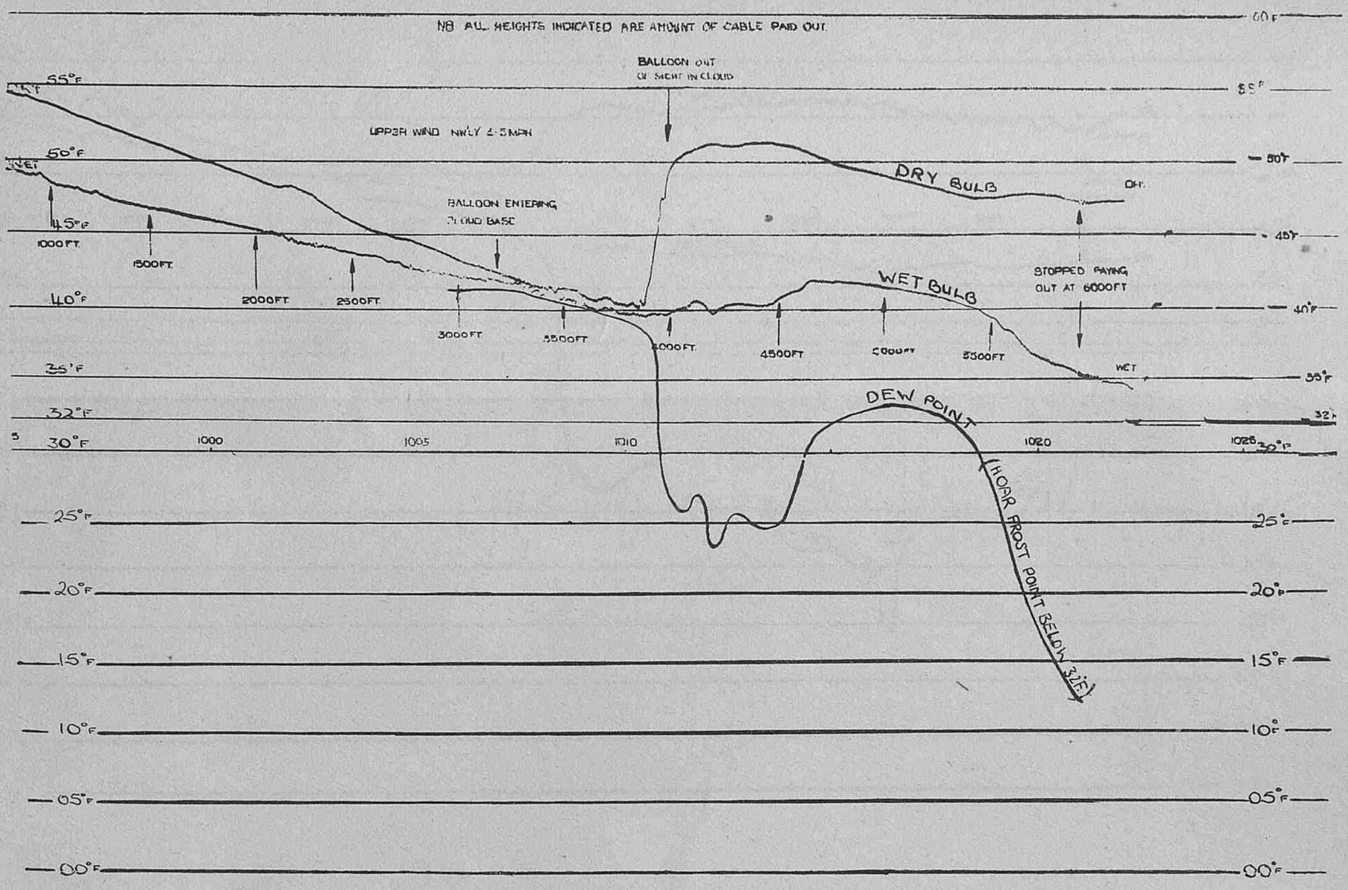


FIG 4.



DATE. 1:10:45. RUN. 1.

TIME: 0941 HRS TO 1023 HRS.

BDE. PSYCHROMETER.

4 CORE TELEPHONE CABLE.

ASCENT AT 200 FEET PER MIN.

WEATHER.

$\frac{10}{10}$ SC. BASE APPROX. 3,500 FT.

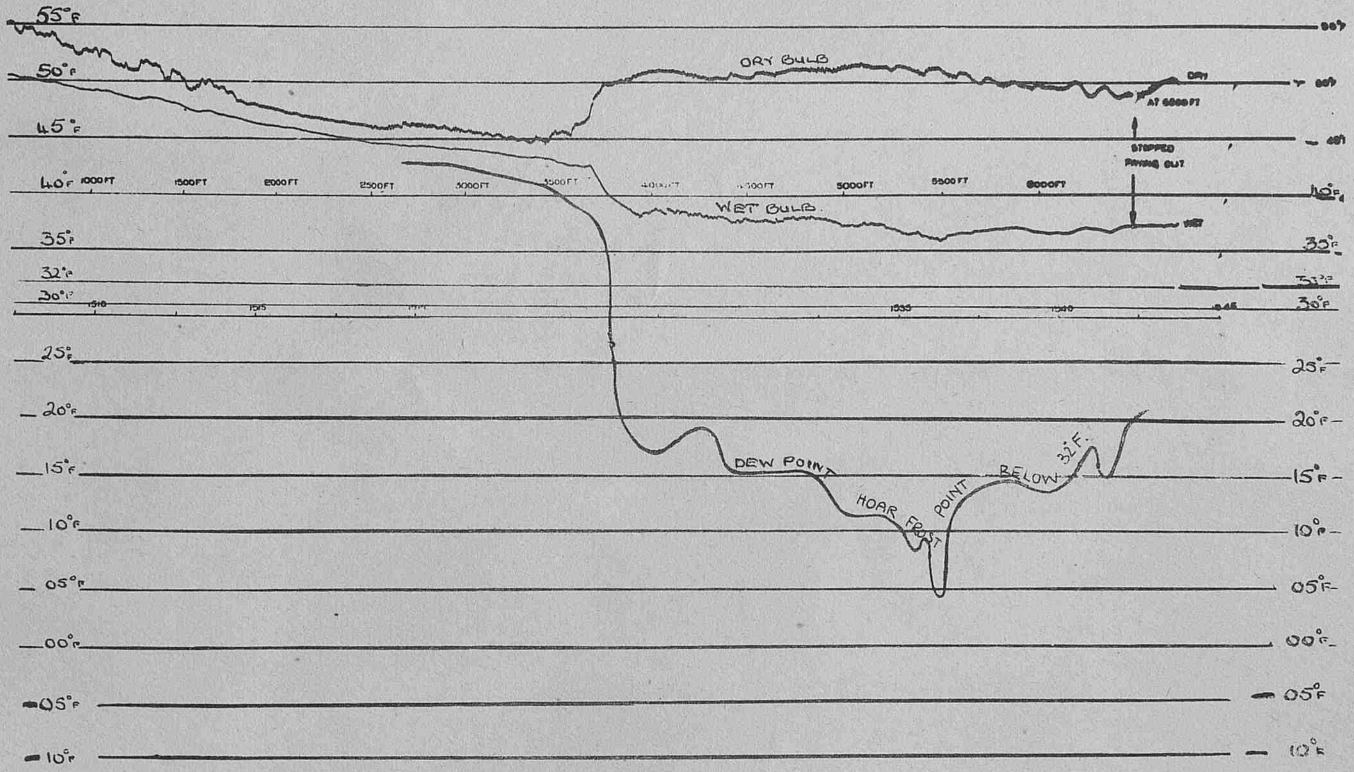
BREAKING AT 1005 TEMPORARILY.

TRACE. F. ST. BELOW, AT END OF TRACE.

SURFACE WIND. CALM TO NE. 1 MPH.

VIS. 1 MILE. IMP. TO 4 MLS.

FIG 5



DATE. 13:10:45. RUN 2

TIME. 1455 TO 1545. GMT.

RANGE SURFACE TO 6500
AT 170 FT. PER MIN.

NO. 1. PSYCHROMETER

NO 4 CORE TELEPHONE CABLE

WEATHER - FINE

SURFACE WIND SE. 5 MPH.

This re
published and
CONFIDENTIALLY.
contents must not be quoted.

16th February, 1946.

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

The Use of an Interferometer to Measure Air Density

by

O. M. ASHFORD

1946 AUG 21 AM 9 51

1. Some factors which will affect the accuracy of the measurements are considered briefly:-

(a) Disturbance of air flow

For maximum compression we have $\Delta p = \frac{1}{2}\rho_0 V^2$
Assuming an adiabatic change $p/\rho^\gamma = p_0/\rho_0^\gamma$

$$\therefore \frac{\Delta\rho}{\rho_0} = \left(1 + \frac{1}{2}\frac{\rho_0}{p_0}V^2\right)^{\frac{1}{\gamma}} - 1$$

This is a convenient form for inserting numerical values, as the velocity of sound is $1.2\sqrt{\frac{p_0}{\rho_0}}$.

At normal temperatures the following results are obtained:-

Speed m. p. h.	$\frac{\Delta\rho}{\rho}$ %
100	0.8
200	3.4
300	7.6
400	13.4

At lower temperatures the changes will be rather greater than those in the table.

In actual practice, the full compression would not be realised, as the instrument would be designed so that the working part (the air gap between the plates) would be as free from obstruction as possible. It is difficult to estimate the compression factor at this stage, but it is likely to be more like 1/5th than unity. Any rapid variations in the air flow would be serious as they would result in an unsteadiness of the fringe system which would adversely affect the resolution.

(b) Resolution

Assuming the resolving power of the optical system to be 1/20th of a fringe and a path length of 10 cm. the smallest change of density which can be measured is:-

0.09% at 1000 mb.
1.05% at 100 mb.
10% at 10 mb.

(Details of the calculation are available.)

(c) Vibration

For aircraft applications, vibration would probably be the most serious problem. Again it is difficult to estimate the actual magnitude of the vibrations, but the order of magnitude can be deduced from R.A.E. Technical Note Vibration No. 13. Without a special anti-vibration mounting the amplitude varies from 8/1000" at 10 cps to 1/1000" at 500 cps. A mounting of the "Lord" type reduces the latter to 2/1,000,000" but increases the former to 16/1000". The wave length of the light is of the order of 2/100,000" and a movement of the interference system of 1/20th of a fringe therefore corresponds to a change in the optical length of 1/1,000,000" . In order to obtain a resolution of 1/20th of a fringe it would be necessary to have the optical system mounted so rigidly that under vibrations of the whole system of the order of 1/1000" the relative position of the plates did not vary so much as 1/1,000,000"

(d) Temperature

The method is sensitive to differential changes of temperature between the two light paths, and it would be necessary to measure the two temperatures fairly accurately. A differential error of 1°C corresponds to an error of some 0.4% in the density measurement. Under conditions of strong radiation, special care would have to be taken.

2. It is clear that the development of a satisfactory interferometer for use on an aircraft is a major undertaking. For radiosonde purposes item (a) does not arise, and item (c) will be much less serious; it is probable that in time an accuracy of 2% at 100 mb. would be achieved. This corresponds to an accuracy of 2 mb. in pressure and 4°C in temperature. When used in conjunction with a good thermometer the instrument would be useful for checking aneroids and other pressure measuring devices, as well as for obtaining direct measurements of density.

AIR MINISTRY. M.R.P. 277.

19th March, 1946.

METEOROLOGICAL RESEARCH COMMITTEE.

~~This report has not yet been published and is to be treated CONFIDENTIALLY. Its contents must not be quoted.~~

The Shape of Raindrops.

by

A. C. Best.

1. Introduction.

It would probably be a matter of some difficulty to compute accurately the shape of a raindrop falling with its terminal velocity. A knowledge of the pressure variation over the surface of the drop would be necessary and these forces would have to be balanced against the gravitational and surface tension forces. A further complexity might arise if circulation occurs in the drop.

The size of a raindrop is usually quoted by specifying the diameter of a sphere having a volume equal to that of the drop even though it is commonly recognised that the shape of the larger raindrops must depart from the spherical. In the following paragraphs, it is proposed to assume that a raindrop is not spherical but has the next simplest shape - that of an ellipsoid of revolution about a vertical minor axis and to compute the constants of the generating ellipse by comparing the terminal velocity of water drops with the terminal velocity of rigid spheres of similar weight.

2. Terminal Velocity of a Rigid Sphere.

The drag F on a sphere of diameter d moving with velocity v through a fluid of density ρ is given by

$$F = \frac{1}{8}\rho v^2 \pi d^2 C \qquad (1)$$

where C is the drag coefficient, a non-dimensional constant depending on Reynolds number. When the sphere is falling with its terminal velocity the drag can be equated to the gravitational force acting on the sphere giving

$$C.\frac{1}{8}\pi\rho v^2 d^2 = \frac{1}{6}\pi(\sigma-\rho)d^3 g \qquad (2)$$

where σ is the density of the sphere. Using the relation

$$R(\text{Reynolds Number}) = \frac{vd}{\eta} \qquad (3)$$

where η is the coefficient of kinematic viscosity of the fluid surrounding the sphere we get

$$v^3 = \frac{4}{3}\frac{(\sigma-\rho)g\eta}{\rho}.\frac{R}{C} \qquad (4)$$

$$d = \frac{R\eta}{v} \qquad (5)$$

The relation between R and C has been given by Castleman (1) in the form of an experimental curve based on the work of a number of investigators. The figures in table 1 have been extracted from that curve.

Table 1.

Drag coefficient for a Sphere (Castleman).

log R	2.0	2.2	2.4	2.6	2.8	3.0	3.2	3.4	3.6	3.8
log C	0.04	$\bar{1}.94$	$\bar{1}.86$	$\bar{1}.77$	$\bar{1}.69$	$\bar{1}.63$	$\bar{1}.58$	$\bar{1}.55$	$\bar{1}.53$	$\bar{1}.53$

Thus, for selected values of R, C can be evaluated and hence v computed from equation (4). The corresponding value of d can then be obtained from (5). From these corresponding values of d and v a curve can be plotted and hence values of v obtained for specified values of d, This procedure has been followed using the following constants (c.g.s. units)

$$\rho = 1.189.10^{-3}, \quad \eta = 0.15, \quad \sigma = 1, \; g = 981.$$

and neglecting ρ in the expression $(\sigma - \rho)$. We thus obtain the values given in table 2 for the terminal velocities of rigid spheres of unit density falling through air at 1000 mbs. pressure and 20°C temperature. It should be added here that, for diameters exceeding 6.5 mm, Castleman's curve was extrapolated. In this region, however, C changes but very slightly with change in R. Also a similar curve has been given by Goldstein (2) which extends well beyond this diameter. Accordingly Castleman's curve was extended using Goldstein's curve as a guide. It was considered preferable to use the former curve rather than the latter since Castleman had used data obtained from the motion of spheres of approximately the same size as raindrops.

3. Terminal Velocity of a Water drop.

This has been dealt with by Sutton (3) who tabulates the terminal velocity of waterdrops of various diameters when falling through air at 1000 mb. pressure and 20°C. These results have been plotted and the terminal velocity of drops with the diameters shown in table 2 have been read from the curve. The values for drops having diameter greater than 7.25 mm. have been extrapolated. It is clear from the shape of the curve - the terminal velocity shows very little change with diameter at these sizes - that the extrapolated values cannot be greatly in error and the reason for assessing the values for such large drops will appear later.

4. The Equation of motion of a Waterdrop falling through air.

Let D and V be the diameter and terminal velocity of a rigid sphere of unit density and let d and v be the diameter of the horizontal equatorial cross section and the terminal velocity of an oblate ellipsoidal waterdrop of unit density with the same volume as the rigid sphere. The gravitational forces on the sphere and the ellipsoid are identical. Thus if C_S and C_E are the drag coefficients for the rigid sphere and the liquid ellipsoid respectively we have from equation 1.

$$C_S V^2 D^2 = C_E v^2 d^2 \qquad (6)$$

where C_S is the drag coefficient for a sphere corresponding to $R = VD/\eta$ and C_E is the drag coefficient of an oblate ellipsoid corresponding to $R = vd/\eta$. For a given value of D we know V, v and C_S. Thus the equation can be solved if C_E can be determined.

5. The Drag Coefficient of an Ellipsoid.

Unfortunately the writer has been unable to find any information about the drag coefficient of an oblate ellipsoid of revolution. It appears, however, that a reasonable approximation can be made along the following lines.

Let the shape of the ellipsoid be specified by the ratio l/d where l is the length of the axis of revolution, assumed parallel to the direction of motion, and d is the diameter perpendicular to this axis. When $l > d$ we have a prolate ellipsoid, when $l = d$ a sphere and when $l < d$ an oblate ellipsoid. It is of course with the last case that we are mainly concerned.

Now C_E will vary both with R and with l/d. It is proposed to assume that the ratio C_E/C_S is independent of R. It is unlikely that this is completely true but some allowance for the approximation can be made by adjusting the value of C_E/C_S for small values of l/d to correspond to fairly high values of R (i.e. to large drops). We can construct an approximate curve giving the relation between l/d and C_E/C_S as follows. When $l = d$ and $l/d = 1$, we obviously have $C_E = C_S$. Thus the curve must pass through the point $l/d=1$, $C_E/C_S=1$.. When l/d is very small the ellipsoid approximates to a circular plate normal to the stream. According to Goldstein (3) (p.65) the drag coefficient of a circular plate is constant at about 1.12. Since C_S varies with R and C_E for l/d 0 is constant the ratio C_E/C_S cannot be independent of R. However the circular plate corresponds to large drops which have a fairly high value for R. Now C_S varies only over the range 0·34 to 0·35 for values of R between 3000 and 12000 (corresponding to rigid spheres of unit density and diameter between 4 and 10 mm. falling through air with their terminal velocity). Accordingly we choose $C_S = 0\ 34$ and so get $C_E/C_S = 1{\cdot}12/0{\cdot}34 = 3{\cdot}29$ when $l/d = 0$. It must also be borne in mind that the flat plate is a limiting case for an ellipsoid and accordingly the l/d/C_E/C_S

curve must approach this point tangentially to the C_E/C_S axis.

A third point on the curve can be obtained from some work by Jones (4). He quotes a case of a prolate ellipsoid of length 24 inches and diameter 6 inches in a wind of 40 ft/sec. The force exerted on the ellipsoid was 0·0211 lbs. Assuming $\eta = 0{\cdot}15$ and $\rho = 1{\cdot}189 .\ 10^{-3}$ and converting the other figures to the C.G.S. system we get

$$C_E = 0{\cdot}0585 \qquad l/d = 4 \qquad R = 1{\cdot}24 .\ 10^5.$$

It was considered that this value of R is too far outside the range of Castleman's curve for C_S for satisfactory extrapolation and a value of 0·44 for C_S was obtained from the curve given by Goldstein and referred to at the end of section 2. In this region of R the drag coefficient of a sphere is not varying rapidly with R though there is of course a sudden marked decrease when R is about $2{\cdot}5 .\ 10^5$. Thus we get the third point

$$l/d = 4 \qquad C_E/C_S = 0{\cdot}0585/0{\cdot}44 = 0{\cdot}133.$$

The three points on the $l/d/C_E/C_S$ curve have been plotted in fig. 1 and a curve fitted to them.

In view of the inadequate data upon which this curve is based two features of it should be emphasised. We are concerned only with that part between l/d = 1 and l/d = 0. Near l/d = 1, i.e. on the part appropriate to the smaller drops, the error cannot be large.

6. The Shape of a Raindrop.

We now return to equation 6. Writing $C_S\ (VD/\eta)$ to denote the drag coefficient of a sphere for $R = VD/\eta$ and a similar expression for C_E this can be written

$$C_S\ (VD/\eta).V^2D^2 = \frac{C_E(vd/\eta)}{C_S(vd/\eta)}.\ C_S\ (vd/\eta).v^2d^2$$

The quotient on the right hand side is the quantity which has been plotted against l/d in fig. 1 and which we have assumed to be independent of R. Accordingly the functional form of this quotient can be omitted. Then, rearranging the equation above we get

$$C_S(VD/\eta).\ \frac{V^2D^2}{v^2} = \frac{C_E}{C_S}.\ C_S\ (vd/\eta).\ d^2 \qquad (7)$$

Now for a given value of D both V and v are known so that the left hand side of this equation can be evaluated (assuming $\eta = 0{\cdot}15$). The equation cannot then be solved directly for d since $\frac{C_E}{C_S}$ depends on l/d and $C_S\ (vd/\eta)$ is known only in the form of a curve. If however we select a number of trial values of d then for each trial value the right hand side of equation 7 can be evaluated for $ld^2 = D^3$ (since the volume of the ellipsoid is equal to the volume of the sphere of diameter D) and hence $l/d = (\frac{D}{d})^3$. Thus C_E/C_S can be obtained from fig. 1 and $C_S(vd/\eta)$ obtained from Castleman's curve. The values of the right hand side of equation 7 are then plotted against the trial values of d and so the final value for d found by comparing the resulting curve with the value given by the left hand side of equation 7. It will be noticed from equation 7 that d depends on the square root of C_E/C_S, thus diminishing the effect of errors in fig. 1.

By this procedure the figures for d and l/d in table 2 have been obtained. The value of l/d is plotted against D in fig. 2. The fact that V and v are identical for D = 1 mm. indicates that up to that size raindrops are spherical, within the limits of experimental error. From D = 1 mm. to D = 4 mm. the value of d is within 20% of that of D but for bigger drops the flattening increases considerably.

/Table 2. ..

Table 2.

D.	V.	w.	d.	$\frac{l}{d} = \frac{D}{d}^{3}$
mm.	cm/sec.	cm/sec.	mm.	
0.5	200	200	-	
1.0	400	400	-	
1.5	568	540	-	
2.0	712	642	2.13	.829
3.0	940	789	3.34	.724
4.0	1123	872	4.7	.616
5.0	1270	907	6.2	.523
6.0	1400	914	7.9	.438
8.0	1600	919	11.6	.328
10.0	1760	920	15.3	.280
12.0	1890	920	19.45	.235

7. The break up of large drops.

Two assumptions have been made in computing the figures in table 2, viz. that a raindrop does actually have the shape of an ellipsoid of revolution and that fig. 1 is reasonably accurate. As an indirect check a consideration of the conditions for a large drop to break up into smaller drops is relevant.

Associated with each liquid drop there is an amount of potential energy represented by the surface tension multiplied by the surface area. In the absence of opposing forces the surface tension would cause the drop to take up the shape which has the least surface area while the volume of the drop remains constant. This shape of minimum surface area is of course a sphere. A large drop falling freely, however, does experience opposing forces in the form of the air pressure and consequently is deformed to a shape which is not spherical.

It seems reasonable to assume that the deformed drop will remain symmetrical about a vertical axis and that hence the components of a disintegrated drop will be approximately equal in volume. Now if one large drop breaks up into n smaller drops the total surface area of the n drops will have a maximum value if they are similar in size. Thus disintegration will not require the permanent addition of surface energy provided the surface area of the ellipsoid is equal to that of at least two similar spheres with the same total volume as the ellipsoid.

If the major and minor axes and eccentricity of the generating ellipse of the ellipsoid are a, b and e respectively, r is the radius of the n spheres, we have the usual formulae as follows

	Surface Area	Volume
Ellipsoid	$2\pi a^2 + \frac{\pi b^2}{e} \log \frac{1+e}{1-e}$	$4/3\pi a^2 b$
n Spheres	$4\pi nr^2$	$4/3\pi nr^3$

and of course $e = \frac{\sqrt{a^2 - b^2}}{a} = \sqrt{1 - l^2/d^2}$

The logarithm in the surface area formula for the ellipsoid is of course a Napierian logarithm.

Equating the two surface areas and the two volumes and converting the Napierian logarithm we easily obtain

$$\frac{2}{(1-e^2)^{\frac{1}{3}}} + \frac{(1-e^2)^{\frac{2}{3}}}{e} \cdot 2{\cdot}3026 \log_{10} \frac{1+e}{1-e} = 4n^{\frac{1}{3}} \qquad (10)$$

Using equation (10) has been computed for various values of e and in fig. 3 n is shown plotted against 1/d ($= \sqrt{1-e^2}$). This curve shows the number of similar spheres which have the same total volume and surface area as the oblate ellipsoid with a shape specified by l/d. It will be noticed that is less than 2 provided l/d is greater than 0·33. From fig. 2 this corresponds to D = 7·9 mm. It does not necessarily follow that a drop of this size, falling through still air with its terminal velocity, must break up. The drop will probably be vibrating. This may provide the trigger impulse necessary to initiate the break up if the drop is unstable and the energy of vibration may also lead to disintegration before the ellipsoidal surface is equal to the area of two spheres, i.e. for smaller values of D. On the other hand when a drop breaks up there is a momentary increase in the surface area and also the components will not be spherical even after their velocity has been reduced to the terminal velocity appropriate to their size. This factor would tend to make the critical diameter greater than 7.9 mm. If this should occur one would expect the very large drop to disintegrate into more than 2 droplets, each of which would be more nearly spherical then half of the original drop would be.

In the published papers dealing with experimental determinations of the terminal velocity of water drops the largest drops quoted have diameters of 3.5 mm. (Schmidt), 5·48 mm.(Flower), 6·0 mm.(Laws) and 6·36 mm.(Lenard). For the present purpose Schmidt's largest diameter is irrelevant since he worked with natural raindrops and so the absence of larger drops is probably due to other causes than instability of the drops. If the other three investigators found it impossible to produce larger drops falling through air at their terminal velocity it may have been due to the difficulty of creating larger drops which are not vibrating with an excessive amplitude. In this connection it is of considerable interest to note that Sutton quotes a terminal velocity for a drop of 7·25 mm.diameter, a figure which is in rather remarkable agreement with the critical diameter of 7·9 mm. deduced above in view of the approximations involved in the present work.

Sutton also mentions the use of special apparatus for making large drops. With this apparatus it was found that when increasingly large water drops were allowed to fall through a height of 40 feet the drops would disintegrate when $\rho d^2 g/T$ reached a value of 15.7. Substituting $\rho = 1$, $g = 981$ and $T = 74$, this gives d = 10.8 mm. Referring to the break up of large drops Sutton says "the drop suddenly flies apart into a circlet of smaller drops possessing quite a remarkable symmetry". It has been seen above that this could be expected on the basis of figures 2 and 3.

8. Energy of Vibration of a Drop.

In the previous section reference has been made to the energy of vibration of a drop. It is of interest to try to compare the magnitude of this with the other energies involved. Rayleigh (5) gives a formula for the energy of vibration of a spherical liquid drop which, for the fundamental mode, reduces to $E = \frac{8\pi A^2 T}{5}$ where E is the energy (potential or kinetic according to the instantaneous state of the drop), A is the maximum displacement and T the surface tension. This may be written $E = 5\,A^2T$. Now consider the energy which must be supplied to a drop having an equivalent diameter of 7·25 mm. in order that it may split in two. From fig. 2 such a drop has shape specified by l/d = 0·358 and this corresponds to n=1·83 (from fig. 3). It is easily verified that the energy required is represented by $\pi . T . D_1^2 \, (2^{1/3} - 1{\cdot}83^{1/3}) = 0{\cdot}037\pi T D_1^2$, where D_1 is the critical value of D at which n=2. Now Rayleigh's formula applies only to spherical drops vibrating with small displacements. It seems reasonable, however, to accept it as giving an answer of the correct order of magnitude for an ellipsoidal drop. And in fact the approximations involved in figures 1 and 2 probably do not justify an attempt at greater accuracy. Equating $5A^2T$ to $0{\cdot}037\pi T\,D_1^2$ and putting $D_1 = 7{\cdot}9$ mm. we get A = 1.2 mm. Thus a vibration with maximum displacement of about 1·2 mm. would provide sufficient energy to cause a falling water drop with equivalent D of 7·25 mm. to break up into two halves.

References.

(1) Castleman, R.A., N.A.C.A. Technical Note No. 231, Washington, 1926.
(2) Goldstein, S. (Editor) Modern Developments in Fluid Dynamics, (Oxford, 1938).
(3) Sutton, O.G., M.R.P.40.
(4) Jones, R., R & M No. 1061, December, 1925.
(5) Rayleigh, Proc. Roy. Soc. XXIX, pp 71-97, App. 2, 1879.

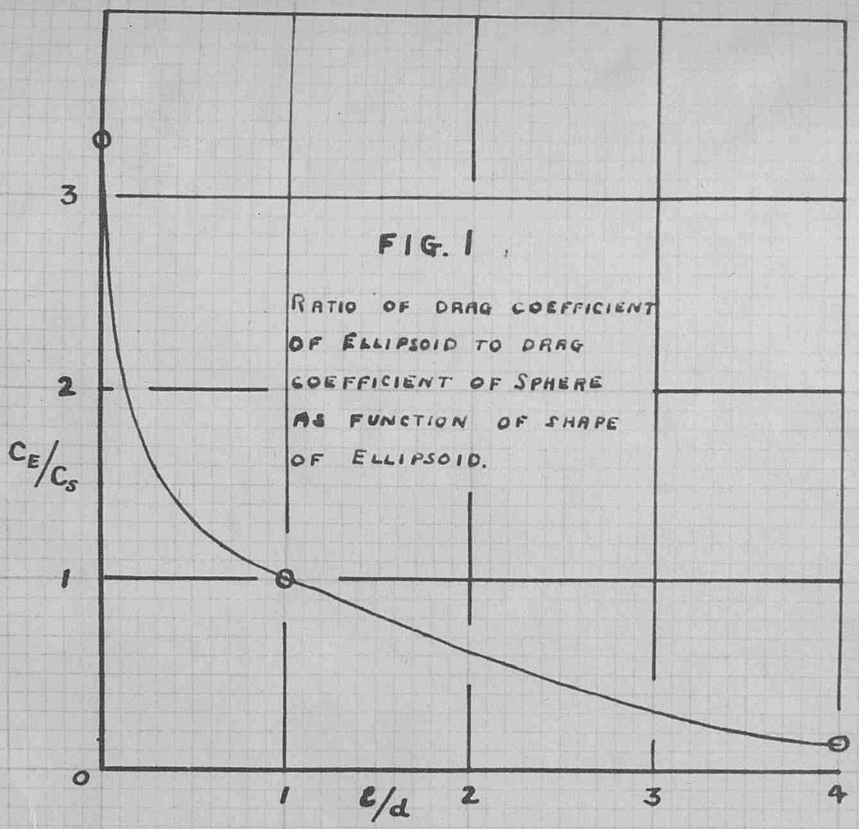


FIG. 1

RATIO OF DRAG COEFFICIENT OF ELLIPSOID TO DRAG COEFFICIENT OF SPHERE AS FUNCTION OF SHAPE OF ELLIPSOID.

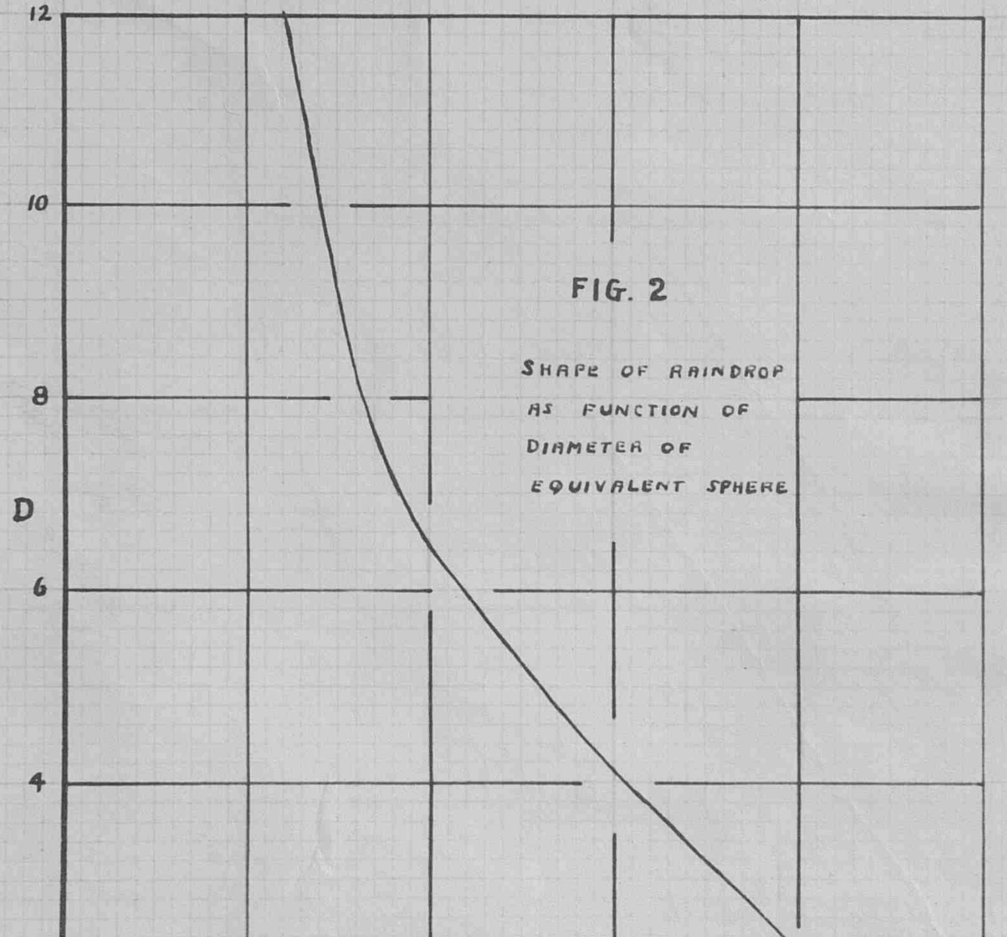


FIG. 2

SHAPE OF RAINDROP AS FUNCTION OF DIAMETER OF EQUIVALENT SPHERE

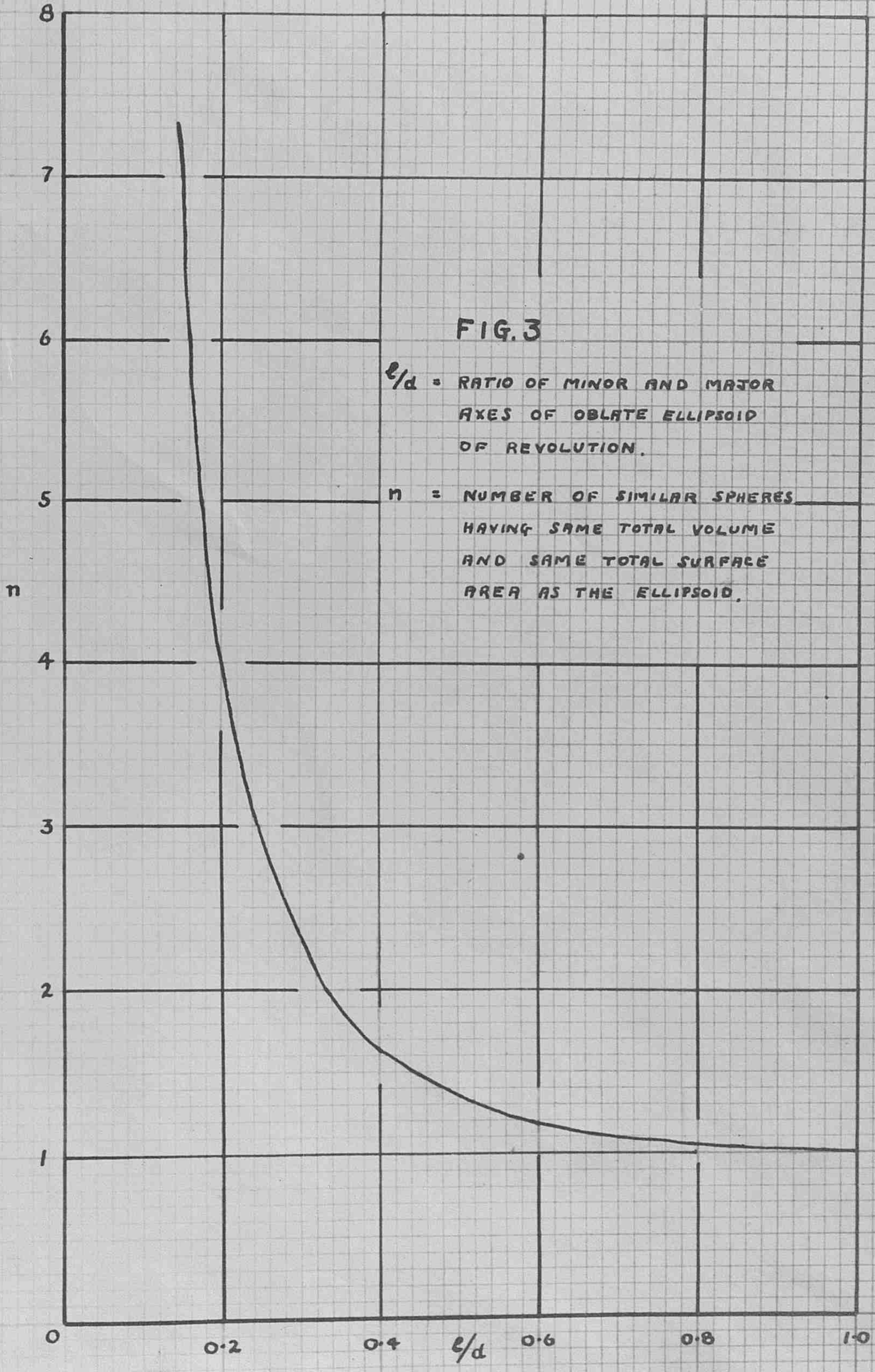


FIG. 3

l/d = RATIO OF MINOR AND MAJOR AXES OF OBLATE ELLIPSOID OF REVOLUTION.

n = NUMBER OF SIMILAR SPHERES HAVING SAME TOTAL VOLUME AND SAME TOTAL SURFACE AREA AS THE ELLIPSOID.